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Part 1

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Edited by

P.C. Thurston, H.R. Williams,
R.H. Sutcliffe and G.M. Stott

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A Message from the Honourable Shelley Martel
Ontario Minister of Northern Development and Mines

The publication of a comprehensive volume on the *Geology of Ontario* is a milestone for the entire Ministry of Northern Development and Mines. I am pleased to be able to participate in this project as Minister of Northern Development and Mines.

Undertaken by the Ontario Geological Survey to commemorate the centennial of the founding of the Ontario Bureau of Mines, *Geology of Ontario* forms part of a comprehensive review and description of the province's geology.

It presents a chronological synthesis of the latest information and concepts on the province's geology. The chapters that follow explain the geological processes that influence our environment. They provide landmark studies of Ontario geology, and we believe they will generate new ideas for mineral exploration.

Seven map themes have been compiled at 1:1 000 000 scale to accompany the volume. These full-colour maps were produced using the latest in computer technology to improve map presentation. Taken together, this volume and its accompanying map folio represent our best geological knowledge in this, our centennial year.

The Ministry is proud of the *Geology of Ontario* project team, and we recognize the painstaking care all contributors put into the production of this volume. Their work will be of immense value to the mineral exploration and development community, both in Canada and abroad, as well as to government geoscientists and academic researchers. The publication will also be of value to secondary school teachers, mineral collectors, professionals in other government departments and members of the media and general public who have an interest in Ontario's geological history and mineral wealth.

As Minister of Northern Development and Mines, I hope *Geology of Ontario* will facilitate a greater understanding of Ontario's geology and a wider appreciation of its importance to the provincial economy.

Foreword

Geology is important to Ontario and all its citizens. As a fundamental control on the economic and social development of Ontario, geology has a tremendous impact. This ranges from influencing the location of hydroelectric power stations and transportation routes to providing the province's many mineral deposits.

Over the last 100 years, the Ontario Geological Survey (OGS) has been the focus of geological understanding for the whole province. From the early days of reconnaissance mapping of bedrock by canoe and portage, to detailed study of mining camps, regional provincial mapping and diversification of activities into glacial geology, geochemistry, geochronology and geophysics, the OGS has led the way to a clearer understanding of the province in which we live.

This volume and map series represent a milestone for the Survey and the province. We have come far, not only with the extent of our geological mapping coverage, but also with our understanding of how all this information interrelates. We have learned that the Ontario we know today has evolved over 3000 million years, through mountain building, volcanism and catastrophe, all of which are preserved in the rock record. What is presented here in the volume and maps is a synthesis of our present state of knowledge.

The Survey, however, is building on its proud past, and growing to meet the challenges of the future. The new Mines and Minerals Research Centre in Sudbury and the continued adaptation and utilization of new technology will facilitate our effectiveness and our leadership role. Mining will continue to be one of the mainstays of the Ontario economy. Nonetheless, economic security and a safe, clean environment need to go hand in hand. We must have a fuller understanding of the landmass to ensure this.

Although "the geology" is often thought of only as rocks, be they ore deposits or granites, "geology", means literally, the study of the Earth. Our atmosphere, our water and the surface we live upon, are all geologically interconnected. Understanding earthquake hazards, the geochemistry of naturally occurring elements and chemicals, assessing land hazards, water movement in the subsurface and mineral resource potential—all have increasingly critical importance to the whole province.

The volume and map series presented here, and the expertise of OGS staff, provide a foundation of geological knowledge to guide future efforts to understand our geological habitat. The success of these efforts will play an essential role over the next century, contributing to Ontario's economic well being, security, health and growth.

V.G. Milne
Director
Geoscience Branch
Ontario Geological Survey

Introduction

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Chapter 1

Geology of Ontario: Introduction

P.C. Thurston

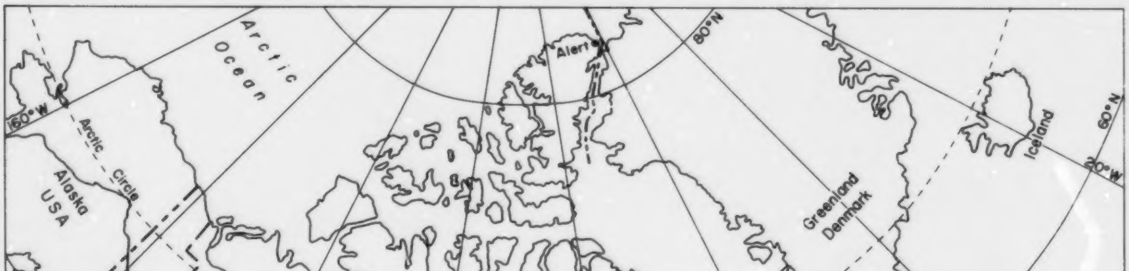
Precambrian Geology Section, Ontario Geological Survey

INTRODUCTION

This volume and the accompanying map series present a synthesis of our knowledge and interpretation of the geology of the province of Ontario. This synthesis gives an account of 1 068 582 km² or 10% of Canada's landmass. Ontario extends about 1400 km from east to west, borders the provinces of Manitoba and Quebec, and spans 1700 km from the southernmost point on the the Great Lakes, north to the shores of James and Hudson bays (Figure 1.1).

This publication celebrates the centenary of the Ontario Bureau of Mines (OBM), established on May 4, 1891, which later became the Ontario Department of Mines (ODM). In 1978, the ODM became known as the Ontario Geological Survey (OGS), now a part of the Mines and Minerals Division of the Ontario Ministry of Northern Development and Mines. This volume and the accompanying maps encompass 100 years of geological studies in Ontario compiled by staff of the Mines and Minerals Division.

The science of geology has undergone recent, fundamental changes which have revolutionized our theoretical framework; for example, the application of global plate tectonic theories to ancient rocks, and breakthroughs in understanding the genesis of gold deposits. Significant geological


Arctic Ocean
Alert
80°N
160°W
Arctic Circle
Alaska USA
Greenland Denmark
Iceland
60°N
20°W
140°W
Whitehorse
Yellowknife
Pacific Ocean
Hudson Bay
Atlantic Ocean
40°W
Manitoba
Edmonton
Calgary
Vancouver
Victoria
Regina
Winnipeg
Ontario
James Bay
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Charlottetown
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Fredericton
Halifax
Montreal
Ottawa
Toronto
Windsor
USA
N
0
800km
120°W
100°W
40°N
40°N
60°W

Figure 1.1. Map of Canada showing the location of Ontario relative to the other provinces (*modified from* Bostock 1970).

and technological advances are incorporated into this compendium, which consists of a map series and an accompanying volume in 2 parts, which explains, in greater detail, the progress these techniques have brought about. These publications provide a data base which will not only be valuable to the mining and exploration industries, but will also be valuable to those concerned with natural resources, environment, power generation, education, earth-science research, prospectors, and municipal and rural planners.

This introduction will highlight briefly the role of geology in everyday life in Ontario and the evolving role of the Ontario Geological Survey. A summary sketch of the physiographic regions of Ontario will be presented as a necessary preface to the various chapters on the geology of the province. There is a summary of the technical highlights of various chapters in order to encourage broad use of the volume.

The Past

The geology of Ontario has influenced the pattern of human settlement since the aboriginal peoples first settled along the numerous rivers and lakes of northern Ontario and developed agricultural communities on arable lands, especially in southern Ontario. European settlers used and expanded the same transportation corridors and agricultural areas. Mineral wealth, first developed by the aboriginal peoples and traded across parts of North America, was reported by early European explorers. Subsequently, waves of prospectors and mine developers established mining communities that influenced the future settlement patterns in northern Ontario. Geology has affected the people of Ontario indirectly by governing the location of early settlements, favourable agricultural land, and sources of hydroelectrical, coal and petroleum energy resources. Thus, our growing knowledge and understanding of the geology has had an effect upon the patterns of socioeconomic development.

For example, bedrock and glacial geology, by controlling the stream gradients, determined the location of water-driven mills around which early communities developed. Later industrial activity concentrated in southwestern Ontario, where local fuel supplies were abundant and the Great Lakes provided a natural shipping corridor to large markets. The Great Lakes provided a seemingly abundant supply of water for the many large communities growing along their margins. The freshwater lakes also served as an industrial coolant, and unfortunately in some cases, the lakes became a repository for municipal and industrial wastes. Yet, a sizeable number of early settlements developed where groundwater was an important resource. Today, almost one third of Ontario's water supply comes from groundwater (S. Singer, Ontario Ministry of the Environment, personal communication, 1991). The future sustainability of this resource will depend upon greater understanding of its geological controls.

Agricultural development thrived in south-central and southwestern Ontario where Paleozoic sedimentary rocks and overlying glacial deposits have contributed to favourable soil conditions. Southeastern Ontario consists of till plains, overlying crystalline bedrock. Here, soil quality is good, but agriculture has in places been hampered by the stony nature of Quaternary glacial deposits. The harsher climate and thin soils of northern Ontario restrict development of agriculture to isolated pockets of arable land. Clay belts deposited by proglacial lakes were covered by lush vegetation and presently form discrete areas of arable land within the rocky, nonagricultural northern part of the province. The northern Ontario clay belts include the areas around Haileybury to New Liskeard, Thunder Bay, Hearst and Dryden.

Glaciation, by redistributing surface material and exposing bedrock, has controlled vegetation density, distribution and variety. The products of the breakdown of bedrock to regolith and eventually soil, create the host for the lush forest regions of Ontario which supply us with wood for pulp and paper products, building supplies and so on.

Geological processes have governed our development to a much greater extent than many realize. Clay and shale beds, gravel and sand deposits, and crushed stone resources throughout Ontario have provided necessary and affordable building supplies for our houses, cities, roads and industrial development. Lignite deposits and peat bogs of the Hudson Bay Lowland provide an enormous, potential fuel source and fertilizer for horticultural applications.

Undisputedly, geological investigations have most benefitted the economic well-being of Ontario by promoting exploration and mining in the Precambrian Shield that so dominates the province. The mining and exploration sectors of the economy are responsible for the settlement of many communities in Ontario. Some of the earliest examples of communities built around mining and exploration in Ontario are the settlements of Cobalt, Haileybury and New Liskeard; they arose from the Cobalt silver camp, which began in the 1880s (Smith 1986). Other Ontario settlements with mining-oriented economies include Red Lake, Manitouwadge, Thunder Bay, Wawa, Marathon, Timmins, Kirkland Lake, Sudbury, Goderich, Hagersville and Caledonia. Pye (this volume) elaborates on the Survey's contribution, over the past 100 years, to the promotion of mining and exploration.

The discovery and extraction of rock and mineral resources has proven to be a diverse and lucrative business (*see* Anders, this volume). Deposits of metals such as copper, zinc, nickel, silver and gold, industrial minerals and rock resources such as salt, nepheline syenite and lime (nonmetals), fossil fuels and building materials such as facing stone, gravel, sand and clay all reflect Ontario's diversity of natural resources. The economic signficance of this is realized by these 1989 production figures (in millions of dollars): metals—$5569; nonmetals—$240; fuels—$84; and structural materials—$1363 (*see* Anders, this volume).

The role of the geologist and the Survey during Ontario's progressive economic growth has been a significant one. The provincial-scale geological maps, the first in 1946 (Satterly 1946) and a second, revised edition in 1958 (Satterly 1958) as well as Ontario Geological Survey reports, provided a data base of geological information as valuable to the exploration geologist and prospector as their hammers. A great deal of Ontario was mapped over the past 50 years in particular; reports and mineral resource circulars began to explain in greater detail our understanding of Ontario's geology. In 1971, the Ontario Department of Mines revised the provincial-scale geological map, incorporating results of reconnaissance mapping of the far north and aeromagnetic surveys of most of the province (Ayres et al. 1971).

The Present

Since 1971, however, the Survey has been implementing the use of many new tools for understanding our geology. New techniques and instrumentation have been implemented in the processes of data collection, storage, manipulation, interpretation and presentation. The Survey continues to build a data base that will serve to encourage the future discovery of new deposits.

Helicopter-supported reconnaissance geological surveys of the far north by the Ontario Department of Mines began in 1966. In co-operation with mapping by the Geological Survey of Canada under the Roads to Resources Program, northern Ontario was almost completely mapped by 1973 at a scale of 1 inch to 4 miles.

The advent of high-precision U-Pb zircon geochronology, made possible largely through co-operative research with the Royal Ontario Museum, has permitted correlation of Precambrian strata and dike sets; established absolute timing of structural and metamorphic events; and has aided in the elucidation of sedimentary rock provenance through detrital zircon studies (*see* Corfu and Davis, this volume).

Computer applications in field-based data collection, storage, retrieval, manipulation and presentation (e.g., the geophysical maps presented with this volume) have been implemented. For example, computer-based processing of aeromagnetic and gravity data permits geological interpretation from coloured, shaded relief plots of these data sets (*see* Maps 2584 to 2599, map case), which enhances the quality of geological interpretation.

Technological advances in geochemical analysis allow more samples to be analyzed with greater speed and accuracy. The better quality data produced by new analytical techniques have: permitted the use of major element geochemistry to support geologic mapping; facilitated the large-scale geochemical classification of metavolcanic rocks (*see* Grunsky et al., this volume); allowed the documentation of igneous and metasomatic processes commonly related to mineralization; and supported studies of sedimentary provenance. The results of trace element and rare earth element geochemistry, and studies of fluid inclusions are presently used to interpret geological settings, predict exploration targets and determine environmental base lines.

Geophysical and geochemical mapping have provided a new dimension to our geological understanding. Techniques for mapping lake sediment geochemistry have been developed and applied in order to provide fundamental environmental information and to characterize the bulk geochemistry of large bedrock units for purposes of mineral exploration (*see* Fortescue, this volume).

Remote sensing techniques have improved our ability to identify structures in bedrock (e.g., Heather et al. 1991). Geographic information systems (GIS) have allowed the development of layered geoscience data bases (e.g., Currie and Ady 1989). Computer-manipulated satellite data have had broad applications in layered data bases and the interpretation of active geological processes such as mass movement and the disruption of drainage patterns.

The Future

The Ontario Geological Survey will continue to stimulate and support the mining and exploration industries operating in Ontario by improving the map data base, by development of a provincially accessible exploration and mineral resource information base to support mineral resource development and land use planning and through further advancement of our geological understanding.

Priorities for the Mines and Minerals Division and the Ontario Geological Survey include working in conjunction with the private sector and universities to develop safe and cost-effective mining methods, cleanup of mining wastes (mine dumps, tailings ponds) and the development of new techniques and instrumentation in mineral exploration, extraction and processing.

The responsibility of the Ontario Geological Survey may, in the future, expand towards new horizons. Expertise from every discipline of geology will be needed to address issues promoting sound use of land, water and geological materials.

Engineering geologists will need to be consulted in rural and urban land management, and to assist in decisions on the location of landfill sites, industrial and/or mine waste sites and potentially, to locate sites for safe nuclear waste disposal. Geological studies in Ontario (Stone 1984; Stone et al. 1984) examining Precambrian bedrock as a potential, safe repository for nuclear waste have led to advances in the understanding of fluid circulation and structures of the bedrock (e.g., Kamineni et al. 1990).

Civil security regarding natural hazards such as earthquakes, cave-ins caused by structural instability, floods and mud slides (e.g., the Scarborough Bluffs; *see* Barnett, this volume) can be predicted, thus, limiting property damage through understanding geological processes.

The interpretation of earthquake seismic data by the Ontario Geological Survey contributes to our understanding of structures and geological processes. Recognition of areas of seismic activity is particularly important in highly populated or industrialized regions, and for safe planning of sensitive facilities such as nuclear power stations and chemical plants.

Evaluation of our renewable and nonrenewable resources will be important in planning for the generations to follow this one. Inventories of metallic mineral deposits are critical to future mapping and exploration priorities. Water resources and their intelligent use require geologically based planning decisions. Environmental base line studies can help to determine where pollution (acid rain) has profoundly affected the natural balance. Advances in analytical geochemistry allow detection and tracing of a variety of elements in groundwater and soils.

Oxygen isotope studies of glacial ice, used to determine the history of climatic changes, have helped us understand the magnitude and importance of global climatic variations. Geological work of this sort bears directly upon recent observations of ozone layer depletion and a possible recent warming trend in the climate (*see* Barnett, this volume).

The Ontario Geological Survey is relocating to Sudbury in mid–1992. Our new research facilities in Sudbury will allow us to advance our knowledge of Ontario geology more efficiently; that is, to apply the science toward sustainable development of the province; and maintain effective interaction with our client groups.

PHYSIOGRAPHY OF ONTARIO

Physiography is the description of surface features of the Earth, emphasizing their mode of origin (Bates and Jackson 1987). The large-scale physiographic subdivisions are described here to introduce terminology used in several chapters. The changing physiography of this province has been controlled, over the past 3 to 4 billion years, by an array of geological processes which include volcanism, sedimentation, plutonism, faulting, mountain building, erosion, burial, uplift, weathering and glaciation. Of the processes that have shaped the surface features of Ontario, the most recent has been glaciation, during which the bedrock surface was carved and scoured; clay, sand and gravel were deposited, and many of the lakes, including the Great Lakes, were shaped into the forms we recognize today. Geological and physiographic distinctions can be made between 2 major physiographic regions of Ontario, the uplifted broad dome of the Canadian Shield and the surrounding flatter lowlands termed the Borderlands (Bostock 1970; Figure 1.2). Compared to the Borderlands, the Canadian Shield is a topographically higher region of more resistant rock, and hence, influences drainage patterns. A system consisting of a myriad of rivers, flows off the Canadian Shield onto the Borderlands, and drains either to the north and east into Hudson and James bays, or south and east into the St. Lawrence River and the Great Lakes.

Transecting both the Borderland and Canadian Shield regions are 5 "arches" or ridges of Precambrian bedrock (*see* Figure 1.2). Although these "arches" no longer form major topographic features, the distribution patterns of Phanerozoic sedimentary rocks around the "arches" indicate that these features had been positive topographic elements controlling deposition of the Phanerozoic sedimentary rocks (*see* Johnson et al., this volume).

The Canadian Shield of Ontario

The dome-like topography of the Canadian Shield, composed of Precambrian bedrock, which includes elements of several tectonic provinces of Precambrian age, forms a peneplained, gently undulating surface which dips shallowly to the north and south (Bally 1989). The shield has been subdivided by Bostock (1970) into the following (*see* Figure 1.2): the Severn upland, which includes the Nipigon plain and the Port Arthur hills; the Abitibi upland, which includes the Cobalt plain and the Penokean hills; and the Laurentian Highlands, which include the Ottawa–Bonnechere Graben.

The Severn and Abitibi uplands (Bostock 1970) have many features in common. Elevations rise gently up from the Borderlands in the north, to a maximum of 450 m above sea level. These uplands form a rocky landscape, scattered with lakes (Figures 1.3a and 1.3b) and large areas which are mantled by deposits from Pleistocene glaciation consisting of the lacustrine clays and former shorelines of proglacial lakes. Landforms include outwash channels, tills and moraines (described *in* Barnett, this volume).

The Nipigon plain, centred on Lake Nipigon, east of Thunder Bay, is a region of deeply incised valleys and spectacular cliffs dominated by gently undulating to flat-lying gabbroic sills overlying Archean and Proterozoic rocks; it is the level top surfaces of these sills that control the topography in this region. The rocks are in places deeply incised by fractures and faults, which produce scattered hills capped by the more resistant sills.

The low hummocky terrain of the Cobalt plain is controlled by flat-lying sedimentary Precambrian bedrock (Figure 1.3c). Ridges produced by diabase intrusions that crosscut the sedimentary rocks interrupt the plain and rise abruptly, 150 to 180 m above the sedimentary units underlying the lowlands. Some older Precambrian rocks are exposed as inliers.

The Port Arthur hills around Thunder Bay consist of ridges and cuestas, produced by the underlying folded sequence of Proterozoic diabase sills and sedimentary rocks. These hills rise steeply along the shores of Lake Superior producing spectacular, rugged scenery (Figure 1.3d). The physiography controlled by these folded Proterozoic rocks can be traced eastward through Michipicoten Island in Lake Superior into the Penokean hills and extend from the North Shore of Lake Huron eastward as far as the Sudbury area.

The southernmost part of the Canadian Shield is called the Laurentian Highlands, comprising a southeast-trending, slightly elevated region underlain by Precambrian bedrock. To the south, the highlands gradually disappear beneath the cover sequences of Paleozoic strata and the southern Borderlands. The Laurentian Highlands extend eastward from Georgian Bay to the Ottawa River and southeastward to the St. Lawrence River. The Ottawa–Bonnechere Graben is a Paleozoic feature cutting the Precambrian bedrock of the Laurentian Highlands along a north-northwesterly trend in the area of the Ottawa River and westward to Lake Nipissing. Paleozoic sedimentary rocks are preserved in the presently shallow, gently sloping valleys of this fault system.

These Precambrian rocks were eroded to a gently undulating land surface, before the deposition of Paleozoic strata, between 250 and 570 million years ago (*see* Johnson et al., this volume). However, a Cretaceous weathering episode (130 to 65 million years ago) in northern Ontario (Machado 1987; *see also* Sage, this volume), and later erosion during glaciation, left behind only scarred Precambrian rocks

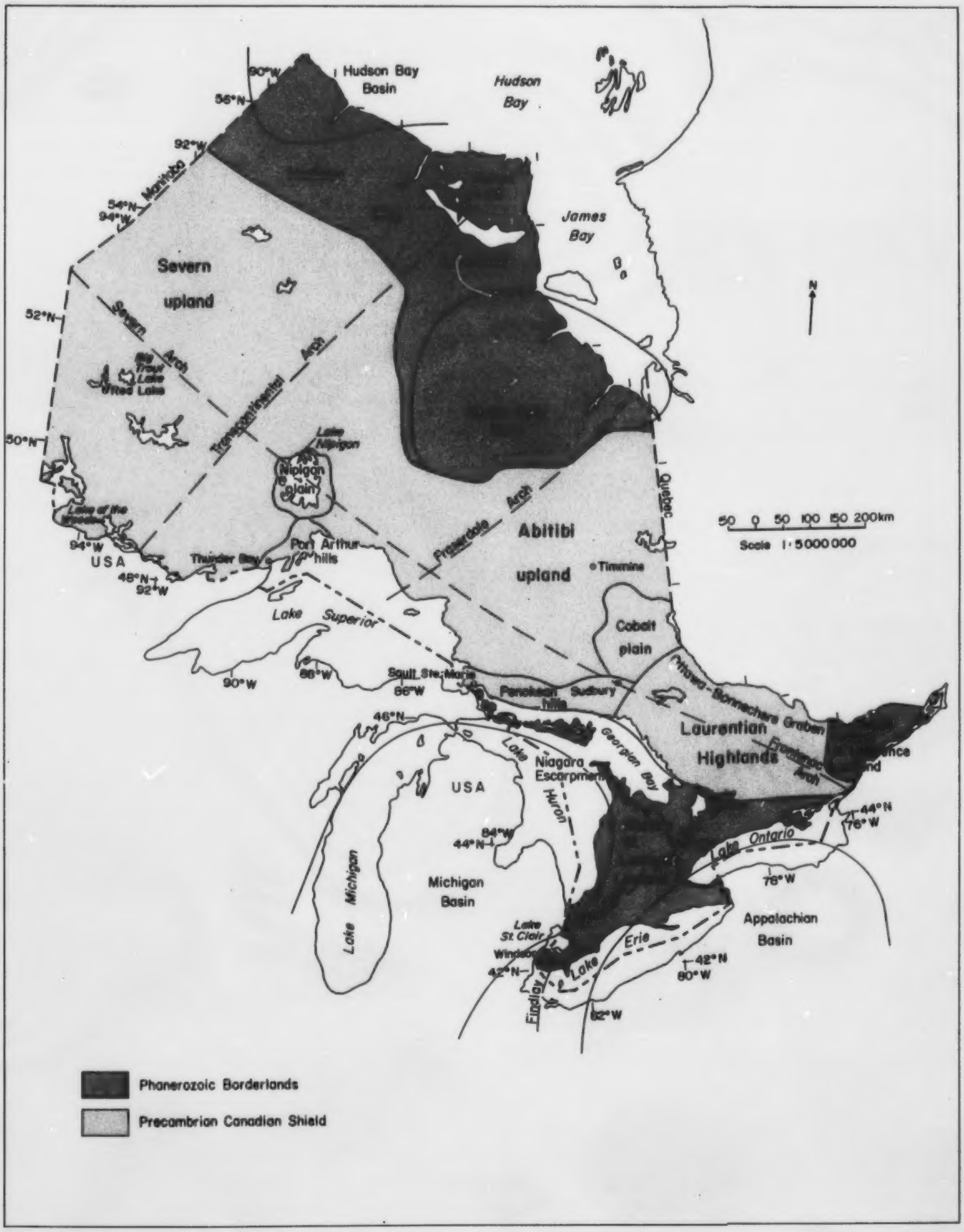
Hudson Bay Basin
Hudson Bay
90°W
56°N
92°W
Manitoba
54°N
94°W
James Bay
Severn upland
52°N
Severn Arch
Big Trout Lake
Sachigo Lake
50°N
Transcontinental Arch
Lake Nipigon
Nipigon plain
Lake of the Woods
94°W
USA
48°N
92°W
Thunder Bay
Port Arthur hills
Fraserdale Arch
Abitibi upland
Timmins
Quebec
Cobalt plain
N
50 0 50 100 150 200km
Scale 1:5000000
Lake Superior
90°W
86°W
Sault Ste. Marie
86°W
Penokean hills
Sudbury
Ottawa-Bonnechere Graben
Laurentian Highlands
Frontenac Arch
St. Lawrence Lowland
46°N
Lake Huron
Georgian Bay
Niagara Escarpment
USA
44°N
76°W
84°W
44°N
Lake Ontario
78°W
Lake Michigan
Michigan Basin
Appalachian Basin
Lake St. Clair
Windsor
Lake Erie
42°N
42°N
80°W
Findlay Arch
82°W
Phanerozoic Borderlands
Precambrian Canadian Shield

Figure 1.2. Map of Ontario showing the physiographic regions *modified from* Bostock (1970).

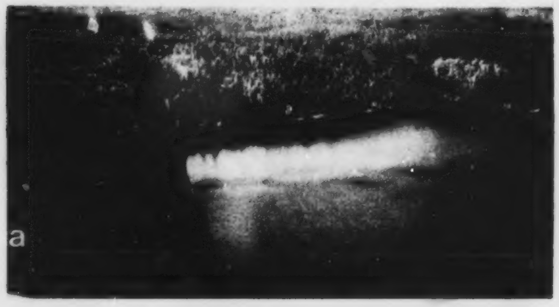

Figure 1.3a. Typical Canadian Shield terrain of the Severn upland. The low relief landscape is commonly rocky and dotted with lakes.

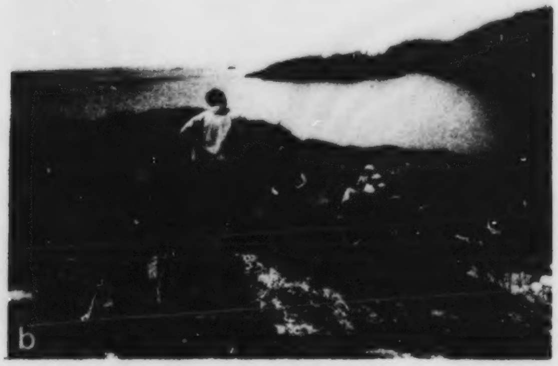

Figure 1.3b. Wave-washed Precambrian rocks of the Canadian Shield on the North Shore of Lake Superior in Pukaskwa National Park.

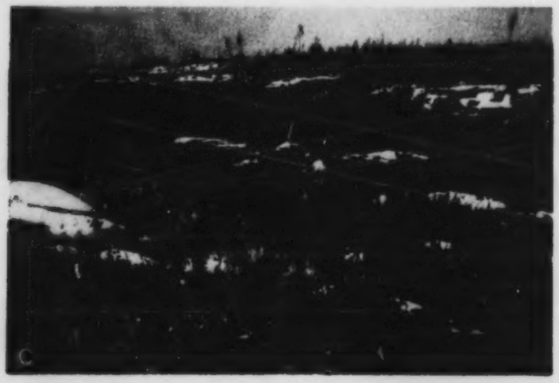

Figure 1.3c. Peneplaned Canadian Shield terrain north of Red Lake. Throughout northern Ontario, burn areas, such as shown here, typically expose abundant, clean outcrop.

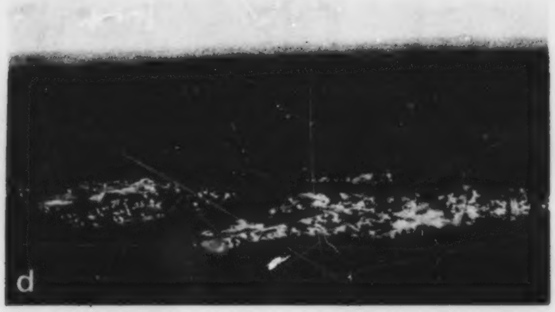

Figure 1.3d. The town of Schreiber, on the North Shore of Lake Superior, is located within the eastern extension of the Port Arthur hills. While terrain here is relatively rugged, ridges and cuestas become more pronounced to the west in the vicinity of Thunder Bay.

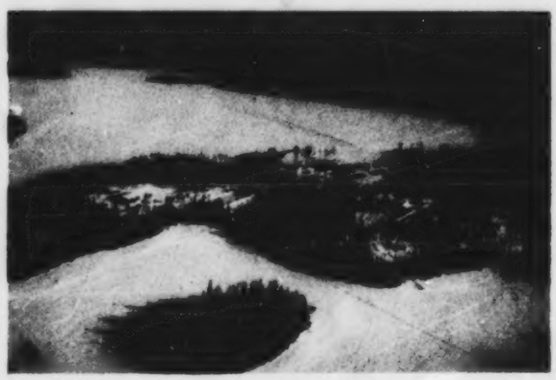

Figure 1.3e. The settlement of Webequie at the eastern margin of the Severn upland adjacent to the Hudson Bay Lowland. The topography of the area is typical of the low relief, relatively flat terrain of the Borderlands.

covered in a few places by flat-lying Paleozoic strata occurring as valley-fill sequences and outliers; for example, the Lake Timiskaming Outlier (Hume 1925; Russell 1981). The face of the Canadian Shield has changed very little since glaciation.

The Borderlands

The Borderlands surrounding the Precambrian Shield form low plains with subdued topography compared to the central highland of the shield. The Borderlands are divided into: the Hudson Bay Lowland, which borders the shield on the north and northeast (Figure 1.3e); and the St. Lawrence Lowlands in southwestern and southeastern Ontario.

On the northern and southern margins of the shield, the Hudson Bay, Moose River and Appalachian sedimentary basins are shallow (a few hundreds of metres) sedimentary basins. The relatively level plains of the lowlands are a function of the flat to gently dipping Paleozoic sedimentary sequences, which are usually blanketed by unconsolidated rock and organic material. Present-day drainage systems flow into these basins and sediments are accumulating in them.

The Hudson Bay Lowland is a low swampy plain, bordering Hudson and James bays (*see* Figures 1.2 and 1.3f). This lowland, with elevations of 60 to 120 m, contains numerous string bogs and shallow lakes (Figures 1.3g and 1.3h). Underlying the beach deposits and peat are flat-lying Paleozoic and Mesozoic sedimentary rocks. The Hudson Bay Lowland is divided into northwestern and southeastern regions by the northeast-trending Transcontinental Arch

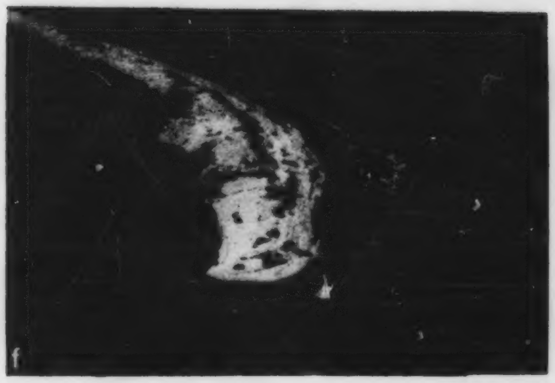

Figure 1.3f. Hudson Bay Lowland raised beach near Paskwachi Point, James Bay. This feature also provides a location for a tagging station for migratory birds (photograph *courtesy of* E.V. Sado).

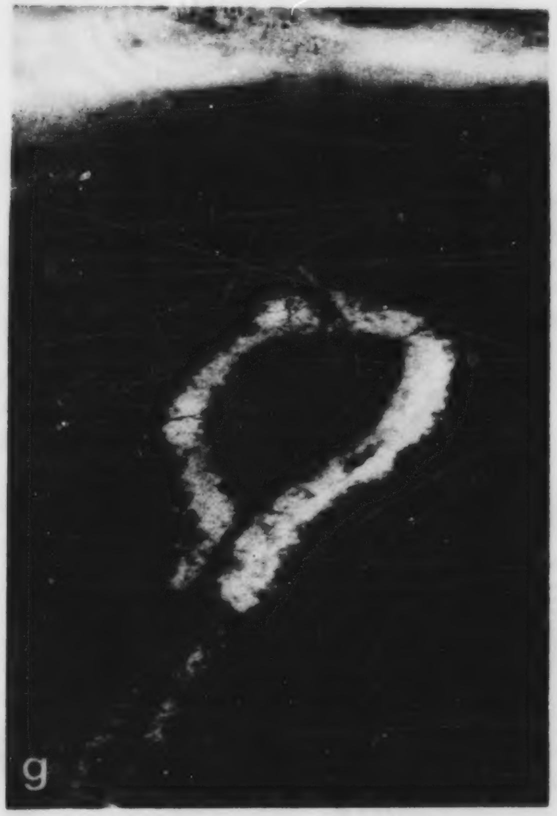

Figure 1.3g. Pothole lake in the Hudson Bay Lowland with non-treed area of muskeg in the background. A beaver dam has blocked drainage from the fen muskeg, creating open water (photograph *courtesy of* E.V. Sado).

(formerly the Cape Henrietta Maria Arch). The Sutton Inlier (Figure 1.3i), a southeast-trending ridge, is a monument of Archean and Proterozoic bedrock, jutting up through the younger Paleozoic deposits. The ridges of the Sutton Inlier interrupt the plain, rising to 150 to 180 m above the lowlands.

Figure 1.3h. Typical muskeg swamp of northern Ontario surrounded by boreal forest. The low swampy plains of the Borderlands account for poor bedrock exposure and make geological mapping difficult.

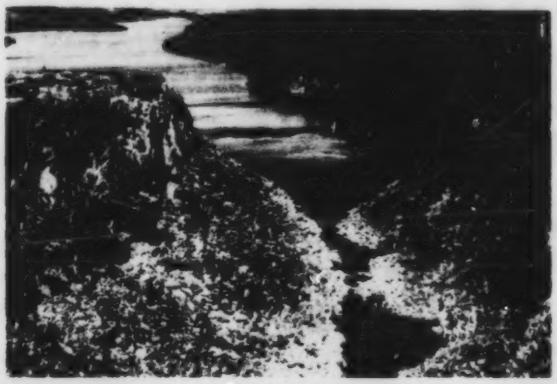

Figure 1.3i. The Sutton Narrows within the north arm of the Sutton Inlier provide a narrow, steep-walled channel interconnecting 2 lakes. Steep cliffs with deeply incised lakes are commonly found within the Sutton Inlier.

The St. Lawrence Lowlands border the Canadian Shield on the southeast and are subdivided into the Western, Central and Eastern parts, of which only the Western and Central parts occur in Ontario (Figure 1.3j). The Western and Central St. Lawrence Lowland physiographic regions correspond to the Michigan and Appalachian basin margins, respectively, and are separated by the southeast-trending Frontenac Arch.

The scenic cliffs of the Niagara Escarpment (*see* Figure 1.2), which stretches across southwestern Ontario, are the product of differential erosion of the north-trending, resistant Paleozoic sedimentary rocks. The ridge has been either enhanced or subdued by the processes of glacial deposition and erosion along its length (Figure 1.3k).

Distributed throughout the physiographic regions are smaller-scale features produced during glaciation. Shallow valleys, depressions, bedrock surfaces and the shorelines of lakes were carved and scoured by the glaciers, while deposits of clay, gravel and boulders were shaped into the eskers, moraines, drumlins, clay belts and numerous other features (*see* Barnett, this volume) found throughout the province.

j

Figure 1.3j. The highly fertile soil of the Chatham flats, east of Lake St. Clair, is primarily recognized for its production of cash crops, such as soybeans. Flat farmland such as this is typical of the topography throughout much of the Western St. Lawrence Lowland in southern Ontario. Photograph *courtesy of* M.D. Johnson, *from* Chapman and Putnam (1984).

k

Figure 1.3k. Niagara Escarpment at "Blue Mountain", south of Collingwood. Silurian dolostones of the Amabel Formation capping the escarpment are poorly exposed here. Ordovician shales of the Blue Mountain, Georgian Bay and Queenston formations underlie the foreground and slope. Photograph *courtesy of* M.B. McLeod, *from* Chapman and Putnam (1984).

Ontario's landscape results from geological processes spanning over 3 billion years of time. Humanity now acts as the strongest physiographic control. In the past 150 years, tunnelling and digging into the Earth's surface, construction of highways, pipelines, railways, water dams, and urban development have significantly modified the face of the province.

NOMENCLATURE AND CONVENTIONS

Terminology used in the volume and map series is based on the following standards: plutonic rock nomenclature of Streckeisen (1976); geochemical classification of volcanic rocks of Jensen (1976), metasedimentary rock nomenclature as summarized by Blackburn et al. (1978); and the metamorphic grade terminology of Winkler (1976). The provisions of the North American Stratigraphic Code (NASCN 1983) are followed wherever possible. The reader should be forewarned that informal stratigraphic nomenclature is used in many parts of the Superior and Grenville provinces because correlation of physically isolated units of similar rock types is uncertain.

The authors of the volume have attempted to distinguish deformation zones in the brittle and ductile regimes through consistent use of terminology listed in the appendix to this chapter. Observations on fault types within the Superior Province are described in the Uchi Subprovince chapter (*see* Stott and Corfu, this volume).

Time subdivisions for the maps and the volume are based on a geochronologic scale for the Phanerozoic and a chronometric scale for the Precambrian. These scales are those adopted for the Decade of North American Geology (DNAG; Palmer 1983) time scale. However, the authors informally subdivide the Archean using elements of the system proposed by Okulitch (1988). A similar subdivision of the Proterozoic by the International Union of Geological Sciences (IUGS; Cowie and Bassett 1989) is followed, with minor modifications. Details are shown on Figure 1.4 and are discussed in the Explanatory Notes for the Map of Tectonic Assemblages (*see* Maps 2575 to 2578, 2583, map case). Subdivisions on Figure 1.4 are designed to include episodes of magmatism and sedimentation.

Most radiometric ages cited were determined at the Jack Satterly Geochronology Laboratory, Royal Ontario Museum, by high-precision U-Pb zircon methods. Where other minerals or other methods are employed, they will be specified. The decay constants used in the calculations are those of Steiger and Jäger (1977). Throughout the volume, the term Ga (billions of years) is used to describe general geochronological relationships and Ma (millions of years) is used where greater precision is intended.

All map themes for the Geology of Ontario have been produced at a scale of 1:1 000 000, using the same planimetric base. Users are reminded that positioning errors involved in creation of any map at this scale mean that the various types of maps cannot be superposed with complete accuracy. The Ontario Geological Survey is aware of a consistent scale difference (ca. 0.1%) between the various types of geophysical maps and the geological maps in the series. This scale difference between geophysical and geological maps is retained since the various geological maps were compiled prior to production of the geophysical maps. Map users should also be aware that the bedrock geology and tectonic assemblage maps were compiled prior to the availability of the shaded relief images used on the geophysical maps. Therefore, geological interpretation using aeromagnetic data was done with contoured magnetic data. Likewise, the gravity data used for interpretation were from the contoured gravity map series of the Geological Survey of Canada (EMR 1974).

STRUCTURE OF THE VOLUME

The various sections of this volume chronologically describe the geology of Ontario. By presenting information in this way, we build up a history of geological events spanning over 3 billion years, from early crustal formation to present-day ongoing surficial processes and neotectonic events. This volume presents new geological models to be tested, while developing new tools for the understanding and synthesis of these models. Large-scale correlations are relatively comprehensive. As selective reporting of small-scale relationships continues, our grasp of the synoptic picture will continue to be strengthened.

The North American Plate was initially subdivided into tectonic provinces[1] (*see* Figure 1.5) based on variation in structural trends (Gill 1949) and the later addition of lithologic and isotopic age criteria (Wilson 1949). In Ontario, these tectonic provinces (Figures 1.5 and 1.6) consist of the Superior Province (Wilson 1949), the Southern Province (Wilson 1939) and the Grenville Province (Wilson 1918). This "provincial" terminology has been employed on the bedrock geology map and in the parts of the various chapters of the volume. The material in the volume demonstrates that the majority of these "provinces" are similar to modern orogens—"linear belts subjected to folding or other deformation" (Bates and Jackson 1987). Indeed, the tectonic provinces referred to above are presently considered orogens or composite orogenic belts. Therefore, the orogenic terminology is introduced in the later stages of the various chapters and on the tectonic assemblage maps. For example, weakly deformed volcanic and sedimentary rocks (Huronian Supergroup) unconformably overlying the Archean in the Lake Huron area were considered part of the Southern Province (Figure 1.7). These 2.4 to 2.2 billion-year-old rocks and their equivalents in the USA (*see* Figure 1.7) were deformed and intruded during the Penokean Orogeny at about 1.8 Ga (Hoffman 1989). Therefore, most of the Southern Province is presently termed the Penokean Orogen (e.g., Hoffman 1989; *see* Figures 1.5 and 1.7), except for rocks of the younger Midcontinent Rift (described by Sutcliffe, this volume). The Superior, Southern and Grenville tectonic provinces form a central stable craton which is flanked by Paleozoic to Mesozoic sedimentary basins (*see* Figure 1.6). These basins were subsequently masked by the products of Quaternary glaciations and minor development of regoliths.

Tectonic Subdivisions

As our mapping coverage has increased, so has our ability to analyze each of the provinces and their characteristic regions. This has led to subdivision of the geological provinces. The most recent of these subdivision schemes for the Archean Superior Province was that of Card and Ciesielski (1986; *see* Figure 1.6); their subprovince subdivisions have been widely accepted. A description of their scheme is presented in the chapter entitled Archean of Ontario: Introduction (*see* Thurston, this volume). The Superior Province is described in chapters based on each of the subprovinces of the Card and Ciesielski (1986) scheme. Proposed revisions to this scheme are summarized in the tectonic summary of the Superior Province (*see* Williams, Stott, Thurston et al., this volume).

Subdivisions of the Proterozoic in Ontario, that is, the Southern and Grenville provinces and the Trans–Hudson Orogen (*see* Figure 1.7) are described later in the volume. The chapter entitled Proterozoic of Ontario: Introduction (*see* Thurston, this volume) briefly introduces the large-scale subdivision of the Proterozoic. The Trans–Hudson Orogen is discussed only in the tectonic summary chapter because it is very poorly exposed in Ontario (*see* Figure 1.6; *see* Williams, Stott, Thurston et al., this volume). The Proterozoic of Ontario is described in chapters on the Huronian Supergroup and related rocks (*see* Bennett et al., this volume), the Sudbury Structure (*see* Dressler et al., this volume) and the Proterozoic of the Lake Superior region (*see* Sutcliffe, this volume). Proterozoic magmatism is discussed in chapters on mafic dikes in the Superior Province (*see* Osmani, this volume) and alkalic complexes and related rocks (*see* Sage, this volume). Subdivisions of the Grenville Province and the geology of the Killarney Magmatic Belt (*see* Figure 13.2 *in* Thurston, this volume) are described in the chapter on the Grenville Province (*see* Easton, this volume).

The Phanerozoic, in contrast to the Archean and Proterozoic of Ontario, is described in slightly different terms. Paleozoic and Mesozoic eras are described (*see* Johnson et al., this volume) in terms of sequence stratigraphy within the sedimentary basins along the margins of this craton (*see* Figure 1.6, Hudson Bay, Moose River, Michigan and Appalachian basins).

Cretaceous erosional events are thought to have been widespread, as demonstrated by the regoliths at Steep Rock Lake (Machado 1987), northern Minnesota (Purucker 1983) and those in the James Bay area (*see* Sage, this volume). Those immediately south of Thunder Bay are described within the Phanerozoic chapter (*see* Johnson et al., this volume).

The chapter on the Quaternary geology of Ontario (*see* Barnett, this volume) describes the deposits and landforms of the Quaternary glacial periods. Glacially derived Quaternary deposits occurring in southern Ontario are described in detail. The description of the Quaternary of northern Ontario is the first provincial-scale synthesis of this area.

Description of Mineral Deposits

The search for mineral deposits, like any other process of geological enquiry, is a complex mixture of the development of theoretical models and the pragmatic process of prospect evaluation and development. Material for both the theoretical and pragmatic aspects of the exploration process is presented. A geological framework for the province of Ontario employing a nongenetic description of crucial

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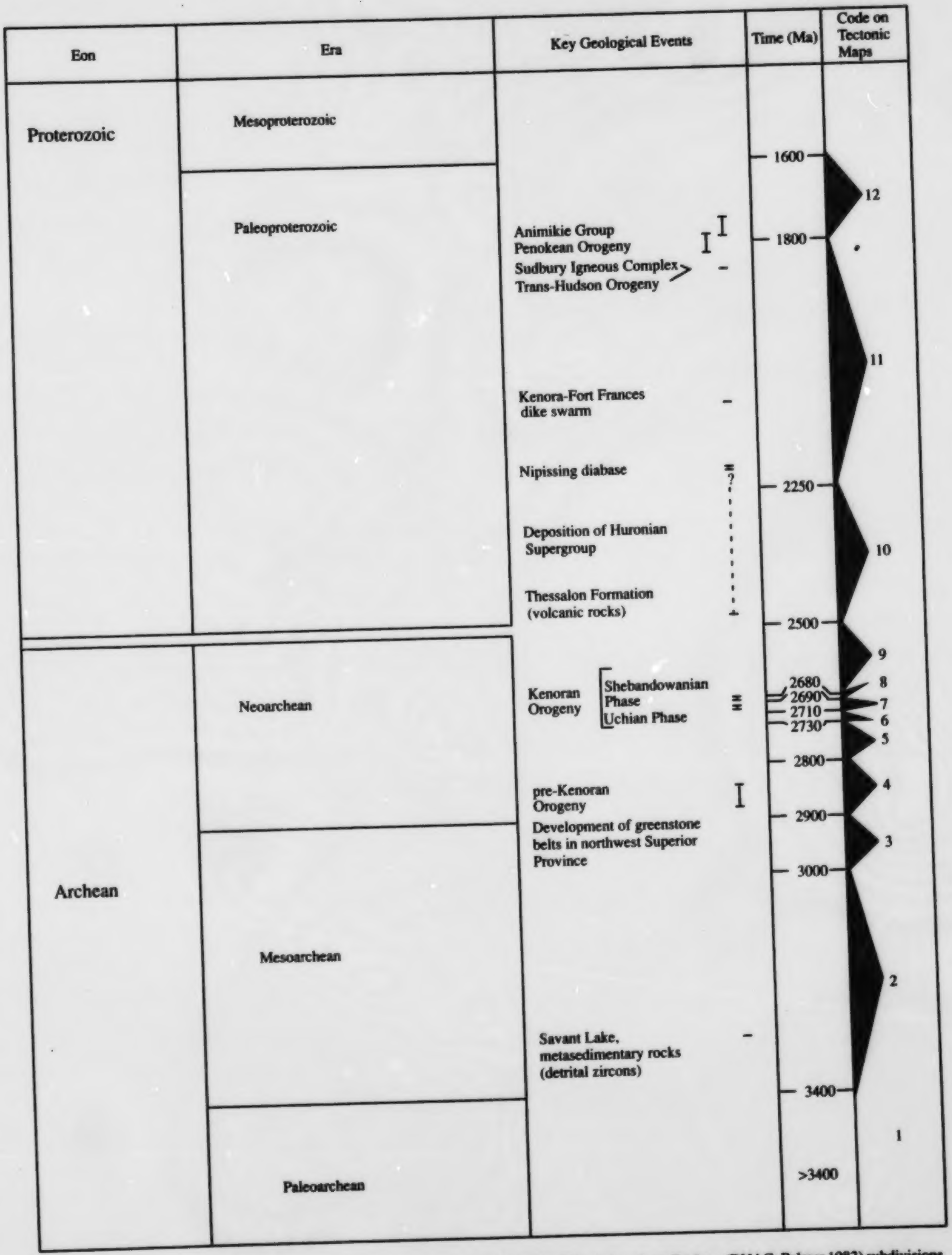
[1] *In Canada, recent literature has tended to use the term "province" as a first order subdivision of cratons. However, North American usage has allowed a degree of interchangeability among the terms "province" (Thomas et al. 1987), "block" (Hatcher et al. 1987) and "craton" (Groves et al. 1987). In this volume, craton refers to crust which has been stable since the end of the Grenville Orogeny at about 1 Ga.*

Eon	Era	Period	Epoch	Key Geological Events	Time (Ma)	Code on Tectonic Maps
Phanerozoic	Cenozoic	Quaternary	Recent/ Holocene	Wisconsinan Glaciation Sangamonian Interglacial Stage		
			Pleistocene	Illinoian Glaciation		
					2	
		Tertiary	Pliocene			
					6	
			Miocene			
					22	
			Oligocene			
					36	
			Eocene			
					58	
			Paleocene			
					63	
	Mesozoic	Cretaceous				
					135	
		Jurassic				23
					205	
		Triassic				
					250	
	Paleozoic	Permian				
					290	22
		Carboniferous		Appalachian Orogeny I		
				Acadian Orogeny I	355	
		Devonian				
					410	
		Silurian				
					438	
		Ordovician		Taconic Orogeny I		21
					510	
		Cambrian				
					570	
Proterozoic	Neoproterozoic					20
					700	
						19
					900	
	Mesoproterozoic			Grenville Orogeny I		18
				Midcontinent Rift, Keweenawan Supergroup I	1120	
						17
				Elzevirian Orogeny I	1220	
						16
				Mackenzie dike swarm I	1270	
						15
					1350	
						14
					1450	
				Killarney Magmatic Belt Anorogenic plutonism Sibley Group I		13
					1600	
	Paleoproterozoic					



Eon	Era	Key Geological Events	Time (Ma)	Code on Tectonic Maps
Proterozoic	Mesoproterozoic		1600	
	Paleoproterozoic			12
		Animikie Group Penokean Orogeny	1800	
		Sudbury Igneous Complex Trans-Hudson Orogeny		
				11
		Kenora-Fort Frances dike swarm		
		Nipissing diabase	2250	
		Deposition of Huronian Supergroup		10
		Thessalon Formation (volcanic rocks)	2500	
Archean	Neoarchean			9
		Kenoran Orogeny: Shebandowanian Phase	2680	8
			2690	7
			2710	6
		Kenoran Orogeny: Uchian Phase	2730	5
			2800	4
		pre-Kenoran Orogeny	2900	3
		Development of greenstone belts in northwest Superior Province	3000	
	Mesoarchean			2
		Savant Lake, metasedimentary rocks (detrital zircons)	3400	
	Paleoarchean		>3400	1

Figure 1.4. The subdivision of geologic time used in this volume employing Decade of North American Geology (DNAG; Palmer 1983) subdivisions in the Precambrian and International Union of Geological Sciences (IUGS; Cowie and Bassett 1989) subdivisions in the Paleozoic and Mesozoic. The right-hand column shows the subdivisions used on the Map of Tectonic Assemblages (*see* Maps 2575 to 2578, map case). The rationale for these subdivisions is discussed in the Explanatory Notes for the Map of Tectonic Assemblages (*see* Map 2583, map case).

observations is also presented so that the information can be of maximum use in exploration, followed by more theoretical syntheses. Throughout the volume, all types of mineral deposits (e.g., metallic minerals, aggregate, industrial minerals, rock resources or petroleum deposits) are described along with their host rocks. Large-scale metallogenic trends are described in separate chapters on metallogeny of the Superior (*see* Fyon, Breaks et al., this volume), Southern (*see* Fyon, Bennett et al., this volume) and Grenville (*see* Easton and Fyon, this volume) provinces. The Grenville Province has not received as much modern exploration as other areas of the Precambrian Shield. Therefore, many of the examples of various types of mineral deposits are prospects or occurrences in contrast to the developed or producing properties used as examples in the other sections of the volume.

Thematic Papers

The 3 thematic papers toward the end of Part 2 of the volume represent an important part of the Ontario Geological Survey program. The papers describe either internal programs of the Survey or programs supported by the OGS which form an important adjunct to the OGS program. The programs described in the thematic papers are the geochemical mapping program of the Geophysics/Geochemistry Section

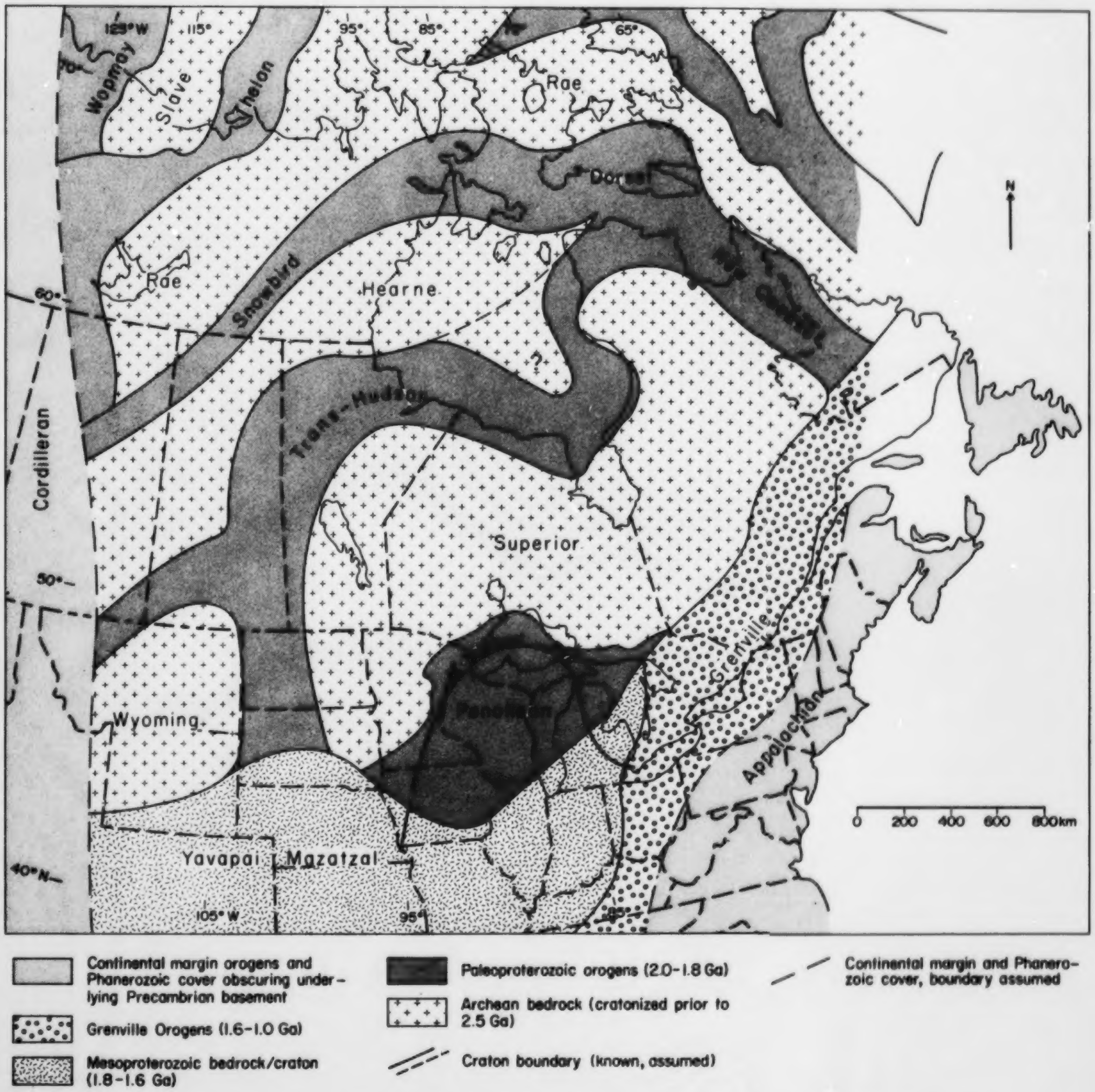

125°W
115°
95°
85°
65°
70°
Wopmay
Slave
Thelon
Rae
Dorset
Rae
Snowbird
Hearne
60°
?
Trans-Hudson
Cordilleran
Superior
50°
Wyoming
Penokean
Grenville
Appalachian
Yavapai
Mazatzal
40°N
105°W
95°
85°
N
0
200
400
600
800km
Continental margin orogens and Phanerozoic cover obscuring underlying Precambrian basement
Grenville Orogens (1.6–1.0 Ga)
Mesoproterozoic bedrock/craton (1.8–1.6 Ga)
Paleoproterozoic orogens (2.0–1.8 Ga)
Archean bedrock (cratonized prior to 2.5 Ga)
Craton boundary (known, assumed)
Continental margin and Phanerozoic cover, boundary assumed

Figure 1.5. The craton of Laurentia showing Archean cratons bordered by Proterozoic orogens. The terminology is taken from Hoffman (1989).

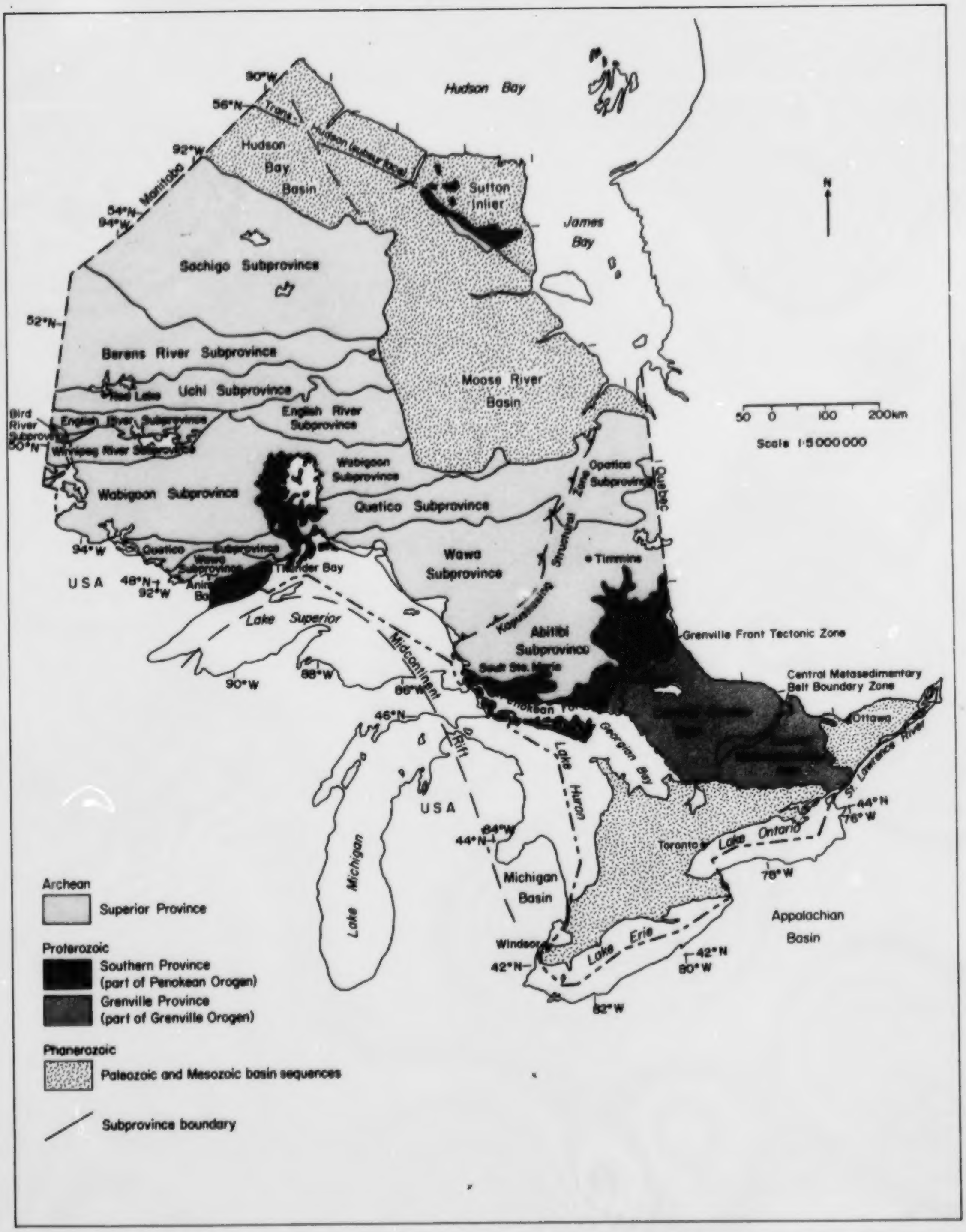
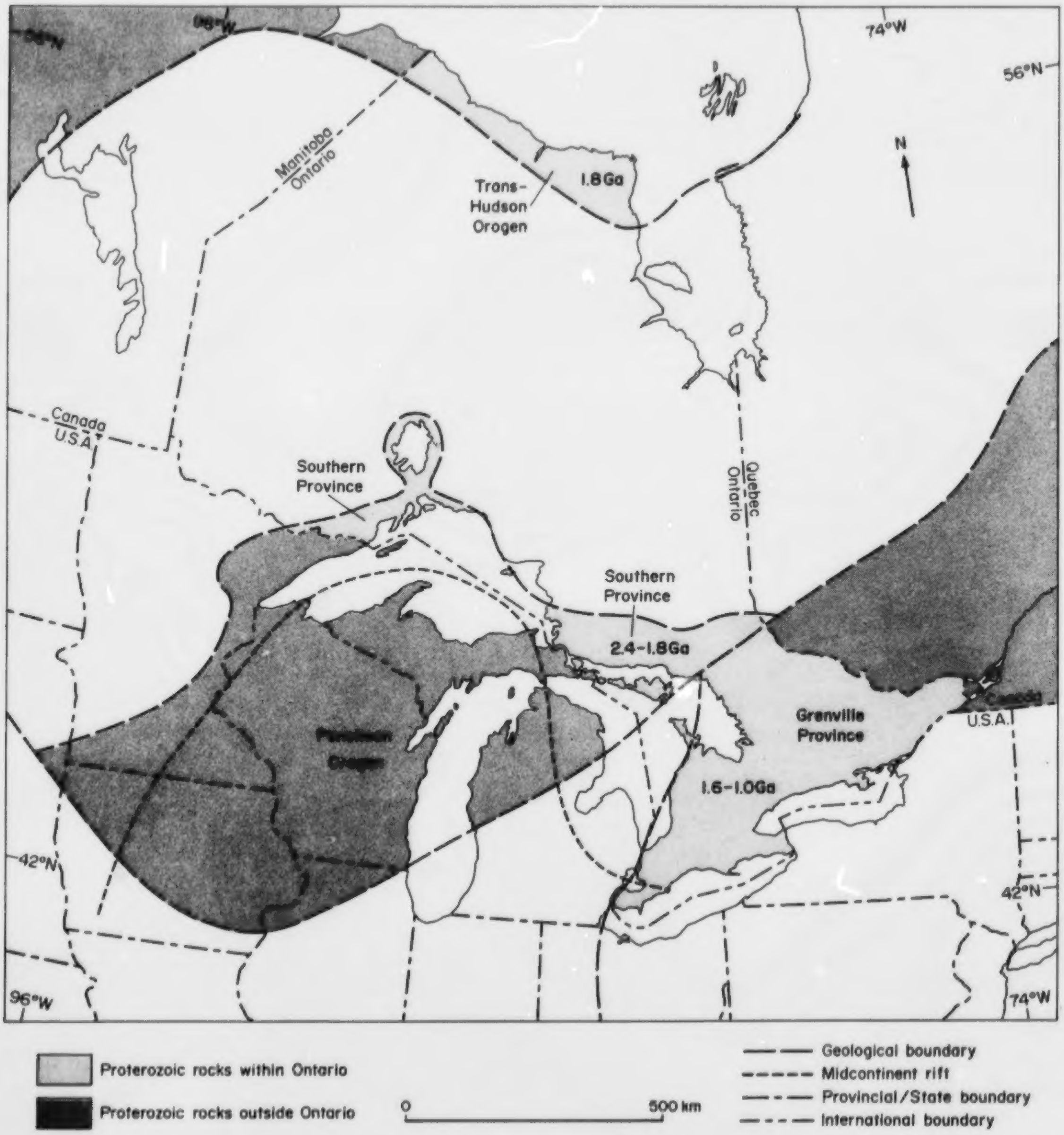

Hudson Bay
90°W
56°N
Trans-Hudson (suture zone)
92°W
Hudson Bay Basin
Sutton Inlier
James Bay
Manitoba
54°N
94°W
Sachigo Subprovince
52°N
Berens River Subprovince
Red Lake
Uchi Subprovince
English River Subprovince
Bird River Subprovince
English River Subprovince
50°N
Winnipeg River Subprovince
Moose River Basin
Wabigoon Subprovince
Wabigoon Subprovince
Quetico Subprovince
Opatica Subprovince
Quebec
Kapuskasing Structural Zone
94°W
Quetico Subprovince
Wawa Subprovince
Thunder Bay
Wawa Subprovince
Timmins
USA
48°N
92°W
Animikie Basin
Lake Superior
Midcontinent Rift
Abitibi Subprovince
Grenville Front Tectonic Zone
Sault Ste. Marie
90°W
88°W
86°W
Penokean Fold Belt
Central Metasedimentary Belt Boundary Zone
Ottawa
46°N
Georgian Bay
Lake Huron
St. Lawrence River
USA
84°W
44°N
44°N
76°W
Toronto
Lake Ontario
78°W
Lake Michigan
Michigan Basin
Appalachian Basin
Windsor
42°N
Lake Erie
42°N
80°W
82°W
N
50 0 100 200km
Scale 1:5000000
Archean
Superior Province
Proterozoic
Southern Province (part of Penokean Orogen)
Grenville Province (part of Grenville Orogen)
Phanerozoic
Paleozoic and Mesozoic basin sequences
Subprovince boundary

Figure 1.6. Major subdivisions of the Superior Province *modified from* Card and Ciesielski (1986) and subdivisions of the Grenville Province *modified from* Wynne-Edwards (1972).

(*see* Fortescue, this volume), the high-precision U-Pb geochronology program conducted jointly by the Ontario Geological Survey and the Jack Satterly Geochronology Laboratory at the Royal Ontario Museum (*see* Corfu and Davis, this volume), and a project to examine the appropriate classification for Archean metavolcanic rocks (*see* Grunsky et al., this volume).

The chapter describing the geochemical mapping program (*see* Fortescue, this volume) is a résumé of a multiyear program employing geochemistry for environmental and exploration purposes. The program has resulted in the development of a multielement mapping method using lake sediments which indexes data to normalized average abundances in the crust (Clarke Index-I Values). The program is traced through its experimental stages to a more complete product. One of the strengths of the technique is a comprehensive data display method which allows rapid interpretation of large amounts of data. Applications of the technique to environmental and exploration problems are many.

The geochronological summary (*see* Corfu and Davis, this volume) synthesizes the fruits of a sixteen-year collaborative effort in the use of high-precision U-Pb geochronology to understanding geological relations within the Precambrian Canadian Shield. The Superior Province



56°N
98°W
74°W
56°N
N
Manitoba
Ontario
Trans-Hudson Orogen
1.8Ga
Canada
U.S.A.
Southern Province
Quebec
Ontario
Southern Province
2.4-1.8Ga
Grenville Province
1.6-1.0Ga
Canada
U.S.A.
42°N
42°N
96°W
74°W
Proterozoic rocks within Ontario
Proterozoic rocks outside Ontario
0
500 km
Geological boundary
Midcontinent rift
Provincial/State boundary
International boundary

Figure 1.7. Major subdivisions of the Proterozoic of central North America.

possesses the most abundant high-precision age determinations of any shield area in the world. The data on the Superior Province are summarized in this chapter. These data have improved our understanding immensely. Correlation of Precambrian stratigraphy, dating of major plutonic suites and dike sets and structural events have contributed immeasurably to our ability to fully analyze the nonfossiliferous Precambrian rocks.

A chapter on the statistical classification of volcanic rocks (*see* Grunsky et al., this volume) reflects work conducted over nearly a decade, initially entirely within the Precambrian Geology Section of the OGS. The project makes use of statistical analysis of major element geochemistry to devise a classification of volcanic rocks which incorporates komatiitic compositions and compensates for the potential mobility of alkali elements in altered Archean metavolcanic rocks. The project has examined variation among all the major elements using multivariate statistics and has: 1) identified the most statistically robust classification digrams; 2) determined which elements are most stable under alteration regimes; and 3) demonstrated the validity in multielement space of the erection of 3 major magma clans for subalkalic volcanic rocks. Additionally, the project has been able to separate the magmatic and hydrothermal imprints present in major element geochemical data for altered metavolcanic rocks.

New data, scientific advances and new interpretations are highlighted below for each of the major sections of the volume.

THE ARCHEAN EON: A STABLE CRATON

Superior Province

Since the mid 1980s, our understanding of the Superior Province has undergone a fundamental revolution through the application of modern structural and tectonic analysis, combined with the applications of geochemistry, geophysics and high-precision U-Pb geochronology. This volume provides a detailed description of advances in our understanding.

The Superior Province is described within a tectonic framework based upon the subprovinces of Card and Ciesielski (1986), using nongenetic terminology derived from the description of modern orogens. This terminology is discussed briefly in the Archean of Ontario: Introduction (*see* Thurston, this volume).

HIGHLIGHTS OF THE SUPERIOR PROVINCE

Here we list some major, newly described relationships within the Superior Province.

- **Assemblages are the fundamental tectonic unit of greenstone belts**

For descriptive purposes, the fundamental tectonic unit within greenstone belts is the assemblage as defined in the chapter entitled Archean of Ontario: Introduction (*see* Thurston, this volume). Development in the volume of the assemblage concept has effectively diminished the importance of the "stratigraphic" model for greenstone belts, which predicted approximately symmetrical patterns of stratigraphy about a central synclinorial axis (Jensen 1985; Nunes and Thurston 1980). Integration of stratigraphy and geochronology, along with the regional mapping of shear zones has shown that greenstone belts more commonly consist of shear zone-bounded, medium-scale units of various ages assembled in a nonsystematic fashion (*see* Jackson and Fyon, this volume; Thurston et al., this volume).

- **Assemblages represent various depositional environments**

Greenstone belts were interpreted to represent mainly cyclic, mafic to felsic volcanism (e.g., Thurston and Fryer 1983; Jensen 1985) partly overlain by Timiskaming-type units of alluvial-fluvial metasedimentary rocks (Turner and Walker 1973) and alkalic metavolcanic rocks (Cooke and Moorhouse 1969). The Superior Province greenstone assemblages are presently considered to represent at least 6 depositional settings (Thurston and Chivers 1990; *see* Jackson and Fyon, this volume), each containing specific rock types and primary structures. The tectonic summary (*see* Williams, Stott, Thurston et al., this volume) defines these 6 settings; specific examples are illustrated in each of the chapters.

- **Mineral deposits and tectonic assemblages**

In Phanerozoic terranes, a temporal, spatial and genetic relationship exists between mineral deposits and specific tectonic settings. Refinements in our understanding of rock associations, structural and metamorphic events and implied tectonic settings has facilitated the application of similar metallogenic scenarios to tectonic settings of the Superior Province; for example, volcanic-associated, massive, base metal sulphide deposits in convergent settings and lode gold deposits in association with strike-slip faulting (*see* Fyon, Breaks et al., this volume).

- **Structural and mineral deposit signatures of cratonization**

The cratonization process includes juxtaposition of contrasting types of subprovinces (granite-greenstone and sedimentary subprovinces), development of late unconformable basins, late granitic plutonism, and transpression or shearing and folding. Archean lode gold deposits, rare metal pegmatites and their parental two mica granites show a spatial association with subprovince and major sequence boundaries. The authors conclude that these deposit types are produced during late-stage crustal thickening and juxtaposition of terranes (*see* Fyon, Breaks et al., this volume).

- **Greenstone belts are tectonic collages**

The collage-like nature of greenstone belts is demonstrated throughout the Superior Province by the juxtaposition of assemblages representing differing depositional settings of various ages. Examples are particularly well demonstrated in the Uchi Subprovince (*see* Stott and Corfu, this volume) and the Abitibi Subprovince (*see* Jackson and Fyon, this volume).

- **Geophysical signature of tectonic units**

Assemblage, terrane and superterrane (subprovince) boundaries are defined, in part, from aeromagnetic, gravity and seismic data. Teleseismic data in northwestern Ontario reveal that the orogenic structural grain of the Superior Province can be traced into the upper mantle, confirming its attachment to the crust since Archean time.

- **Integrated structural history and deformation style for sedimentary and granite-greenstone subprovinces.**

The structural history and deformation style of granite-greenstone and sedimentary subprovinces provide evidence for a generalized scenario of structural development throughout the Superior Province (*see* Williams, this volume; *see also* Stott and Corfu, this volume). For the first time, evidence for an early thrusting is described for all granite-greenstone subprovinces. Relatively late formation and juxtaposition of the sedimentary subprovinces against granite-greenstone subprovinces is explained (*see* Williams, Stott, Thurston et al., this volume).

- **Sedimentary subprovinces are accretionary prisms**

Recent synthesis of structural, sedimentological and geochronological data suggests that the Quetico Subprovince is an accretionary prism on the margin of the older arcs of the Wabigoon Subprovince (Percival and Williams 1989; Williams 1990). Similar conclusions are drawn here for the origin of the English River Subprovince (*see* Breaks, this volume) and the Pontiac Subprovince (*see* Jackson and Fyon, this volume) with respect to their adjacent granite-greenstone subprovinces.

- **Plutonic subprovinces are mid-crustal equivalents of granite-greenstone subprovinces**

Plutonic subprovinces have been considered either microcontinental slivers forming basement to granite-greenstone subprovinces (Hoffman 1989) or deeply eroded mid-crustal equivalents to granite-greenstone subprovinces (Ayres 1969). An example of this controversial relationship is that between the Winnipeg River (*see* Beakhouse, this volume) and the Wabigoon (*see* Blackburn et al., this volume) subprovinces. In another example, the authors conclude from geothermobarometric and stratigraphic evidence that the Berens River Subprovince of Card and Ciesielski (1986) is a more deeply eroded equivalent of the granite-greenstone regions to the north and south (*see* Thurston et al., this volume).

- **New subdivisions of the northern Superior Province**

Prior to our compilation, the northwestern Superior Province was subdivided by Card and Ciesielski (1986) into the Berens River, Sachigo and Uchi subprovinces. This region has been subdivided into 5 newly defined terranes, characterized by distinct lithostratigraphic associations of supracrustal rocks, structural styles and age ranges of supracrustal and intrusive rocks.

- **Comparison with modern orogens**

Rocks and structures throughout the Superior Province are described using nongenetic tectonic terminology; that is, terminology which is not applied exclusively to one type of tectonic hypothesis, such as plate tectonics. However, discussion is also entered into by various authors, comparing the assembly of the Superior Province to modern orogens. The variability of rock types, structural style and tectonic environment suggests that a process of accretionary tectonics best explains the observed patterns. In accretionary tectonics, island arcs collide with older cratonized fragments and develop adjacent accretionary prisms on their outboard edges. Evidence is presented for such processes in the later stages of all chapters discussing the Superior Province. The variations seen on the scale of the province as a whole are presented within the tectonic summary chapter (*see* Williams, Stott, Thurston et al., this volume).

THE PROTEROZOIC EON: OROGENS

Proterozoic orogens occur along the margins of the major blocks of Archean crust of North America (*see* Figure 1.5). The Archean Superior Province of Ontario is bordered by the Mesoproterozoic Trans–Hudson, Penokean and Neoproterozoic Grenville orogens (*see* Figure 1.7). The Proterozoic orogens of Ontario are characterized by a variety of environments: foredeeps, passive margins, foreland thrust-fold belts, hinterlands with basement reactivation, thrusts, transcurrent shears and major rifts.

The Sutton Inlier in the Hudson Bay area (*see* Figures 1.2 and 1.6) has been correlated with the Trans–Hudson Orogen (Hoffman 1989), which is otherwise not exposed in Ontario.

Rocks of the Trans–Hudson Orogen separate the Hearne (of the former Churchill) Province from the Superior Province in northwestern Manitoba. The orogen represents a 1.8 billion-year-old collision zone between these 2 Archean blocks and extends northeast from the Manitoba–Saskatchewan border area through the coastal zone and islands of Hudson Bay, with inliers onshore beneath the Phanerozoic Hudson Bay Basin along the south coast of Hudson Bay in Ontario.

Penokean Orogen (Southern Province)

The Penokean Orogen is a fold belt that deformed various Archean and Proterozoic rocks about 1.8 Ga. The folded rocks include the 2.4 billion-year-old Huronian Supergroup—a succession of rift volcanic rocks and passive continental margin clastic rocks and the synorogenic foredeep (Hoffman 1987) sedimentary rocks of the 2.2 billion-year-old Animikie Group (*see* Figure 1.6). The Huronian Supergroup consists of 4 unconformity bounded siliciclastic groups (Zolnai et al. 1984) which taper northward from a maximum thickness of 12 km. The Huronian Supergroup

(Cobalt Embayment, Figure 1.6) was deposited in an interval *circa* 2.48 to 2.22 Ga.

The eastern end of the Huronian Supergroup is cut by rocks of the Sudbury Igneous Complex (*see* Dressler et al., this volume), an 1850 million-year-old elliptical, norite-gabbro-granophyre intrusion emplaced into brecciated country rock. This structure, together with the rocks of the Whitewater Group, form the Sudbury Basin. During the 1.86 billion-year-old Penokean Orogeny (Hoffman 1989), the southern extremity of the Penokean Fold Belt (*see* Figure 1.6) was shortened and metamorphosed during the docking of the Pembine–Wausau Terrane against the Superior Province along the Niagara Suture Zone.

Little deformed sedimentary and volcanic cover sequences began to overlap the accreted orogens starting at 1.8 Ga. Proterozoic anorogenic plutons cut the craton mainly south and west of Ontario. The cover sequence in Ontario is represented by the ~1.3 billion-year-old Sibley Group in the Lake Superior area (*see* Sutcliffe, this volume), composed of quartz arenite, argillaceous dolostone and mudstones deposited in a playa lake environment.

The 1.1 billion-year-old Midcontinent Rift is a major structure centred beneath Lake Superior (*see* Figures 1.5 and 1.7). Rocks of the Keweenawan Supergroup, composed of subaerial tholeiitic flows, felsic volcanic rocks and minor sediments, are unconformably overlain by coarse clastic rocks and quartz arenite. The rift sequence was deposited in asymmetric grabens.

The Superior Province underwent fracturing related to some of the major tectonic events occurring around it during the Proterozoic and Phanerozoic. These fractures have been filled by 9 separate diabase dike swarms ranging from 2.5 to 1.1 Ga in age, excluding swarms cutting the Grenville Province (*see* Osmani, this volume). Major shear zones transecting the craton are marked by alkalic complexes which have been dated to span the following age ranges: 2.7 to 2.5, 1.70 to 1.85, 1.25 to 0.9 Ga and 650 to 450 Ma (*see* Sage, this volume).

KILLARNEY MAGMATIC BELT

The Killarney Magmatic Belt (van Breemen and Davidson 1988) is a region of anorogenic, 1.7 to 1.4 billion-year-old granitoid plutons and related volcanic rocks lying between the Grenville Province and the Superior Province, south of Sudbury. This block, representing a terrane attached to the Superior Province at some time before the ~1 billion-year-old Grenville Orogeny, is described by Easton (this volume).

HIGHLIGHTS OF THE SOUTHERN PROVINCE

This section of the volume builds upon the recent syntheses of Proterozoic geology of the Great Lakes area (Medaris 1981; Medaris et al. 1983; Hoffman 1988). Listed below are the highlights and significant advances in understanding of the Proterozoic Southern Province, as are discussed in the chapters of this volume.

- **Updated comprehensive discussion of the stratigraphy of Huronian rocks**

Sedimentary rocks of the Huronian Supergroup, along the North Shore of Lake Huron, form 4 unconformity bounded siliciclastic wedges. The Livingstone Creek Formation is newly recognized as a pre–Thessalon unit of quartz-rich arkose and conglomerate (*see* Bennett et al., this volume), thus revising previous views of stratigraphy at the western end of the Huronian outcrop belt.

- **Tectonic subdivision of the stratigraphy of Huronian rocks**

The Huronian rocks have been subdivided into 3 assemblages, with the Elliot Lake Group now split into 2 assemblages, a basal Huronian assemblage attributed to a rift basin environment and a lower Huronian assemblage reflecting a passive continental margin setting (*see* Bennett et al., this volume).

- **Integrated tectonic model for the Huronian rocks of Ontario**

A tectonic model incorporating recent absolute age determinations and revisions to stratigraphy described above provide an updated framework for future work (*see* Bennett et al., this volume).

- **Improved constraints on regional correlation in the Proterozoic of the Great Lakes region**

Discussion of the Proterozoic of the Lake Superior area incorporates recent age determinations on the Lower Keweenawan volcanic rocks (Davis and Sutcliffe 1985; Palmer and Davis 1987), which provide constraints on Keweenawan volcanism and sedimentation (*see* Sutcliffe, this volume).

- **Large-scale plateau volcanism in the Midcontinent Rift**

The volume of basaltic volcanism in the Midcontinent Rift as seen through the integration of reflection seismic data (Cannon et al. 1989) with onshore data shows the Midcontinent Rift to be one of the largest examples of plateau volcanism on earth (*see* Sutcliffe, this volume).

- **Diabase dikes cutting the Superior Province**

The chapter on Proterozoic diabase dikes incorporates a recent flurry of U-Pb dates on diabase dikes, which allows for the refinement of tectonic contexts for each of the 9 major swarms of dikes cutting the Superior and Southern provinces (*see* Osmani, this volume).

- **Alkalic intrusions and kimberlites in the Superior Province**

Proterozoic alkalic intrusions are summarized in terms of petrology, age and form of intrusion (*see* Sage, this volume). Intrusive episodes are centred on 2.7 to 2.5, 1.85 to 1.7, 1.25 to 0.9 Ga and 650 to 450 Ma and *circa* 150 Ma. Definite spatial associations with major crustal structures are described in the chapter. The major structures with spatially associated alkalic intrusions include the Kapuskasing Structural Zone (*see* Williams, Stott, Heather et al., this volume), the Trans–Superior Tectonic Zone (*see* Sutcliffe, this volume) and the

Ottawa–Bonnechere Graben (*see* Sage, this volume; *see also* Johnson et al., this volume). There is a brief summary on the location and petrology of kimberlites cutting the Superior Province (*see* Sage, this volume).

- **Sudbury regional summary**

The enigmatic Sudbury Structure and its copper-nickel ore deposits are summarized along with recent studies on the genesis of the deposits. This includes recent work on the development of various breccias and Sm-Nd isotopic work on sources of the Sudbury host rocks (*see* Dressler et al., this volume).

Grenville Province

The Grenville Province comprises nearly 1 000 000 km² of crystalline rocks forming the youngest part of the Canadian Shield, cratonized around 1 Ga. The Grenville Province was first distinguished from neighbouring Precambrian provinces by structural differences. It forms a 300 to 500 km wide by 2000 km long, northeast-trending belt along the southern margin of the craton, and extends from southwestern Ontario to the shores of Labrador (*see* Figure 1.5). The Grenville Province extends beneath Phanerozoic cover through most of southern Ontario. Outliers of possible Grenville Province rocks are found, for example, within the Appalachian Orogen to the southeast near New York City and Philadelphia.

Within Ontario, the Grenville Province encompasses an area of about 70 000 km² lying southeast of Sudbury, through to the St. Lawrence River (*see* Figure 1.6). The province is characterized by medium- to high-grade metamorphism and ductile deformation; northwest-directed tectonic transport; southeastward younging; thrusting and reworking of older crust. Subdivisions of the province (*see* Figure 1.6), based on Wynne-Edwards (1972), from northwest to southeast are: 1) the Grenville Front Tectonic Zone; 2) the Central Gneiss Belt; and 3) the Central Metasedimentary Belt. The Grenville Front Tectonic Zone consists of Superior and Southern province units bearing Grenville age structural, magmatic and metamorphic (greenschist to granulite grade) overprints. The Central Gneiss Belt, composed of amphibolite to granulite grade quartzofeldspathic and pelitic gneisses, is divided into 4 terranes of Mesoproterozoic age. The Central Metasedimentary Belt consists of belts of juvenile Mesoproterozoic crust, largely supracrustal rocks, generally at amphibolite to granulite grade. The Central Metasedimentary Belt has been subdivided into 5 terranes.

Grenville Province geology is described in a single comprehensive synthesis (*see* Easton, this volume), which integrates new data from a number of fields to support substantial revisions to the interpretation of this complex province. The first gold discovery in Ontario was in the Grenville Province. A variety of deposit types, including gold, various industrial minerals, building stone, iron, lead, zinc and uranium occur in the Grenville Province. Grenville Province metallogeny is examined in a separate chapter (*see* Easton and Fyon, this volume).

HIGHLIGHTS OF THE GRENVILLE PROVINCE

Some of the major results of this synthesis are listed, as follows, in summary form. Details are available in the applicable chapters.

- **Modified tectonic subdivisions are erected in the Central Gneiss Belt and Central Metasedimentary Belt**

Modifications to previous tectonic subdivisions of the Central Gneiss Belt and Central Metasedimentary Belt are offered. The Central Gneiss Belt is subdivided into the Tomiko, Nipissing, Algonquin and Parry Sound terranes. The Central Metasedimentary Belt is subdivided into 5 terranes: the Elzevir, Mazinaw, Sharbot Lake, Frontenac and Bancroft terranes.

- **Spatial and temporal patterns within the Central Metasedimentary Belt**

Patterns of spatial and temporal variation of volcanism and plutonism within the Central Metasedimentary Belt are suggested.

- **Two orogenic events within the Central Metasedimentary Belt**

There are 2 orogenic events within the Central Metasedimentary Belt, an earlier Elzevirian Orogeny (ca. 1.18 Ga) and a later Ottawan Orogeny (ca. 1.1 Ga). The earlier event was probably the more significant of the 2 events in establishing the character of the Central Metasedimentary Belt as it is presently preserved. There is evidence that both of these events also affected the Central Gneiss Belt.

- **Linking of terranes within the Central Metasedimentary Belt**

The Central Metasedimentary Belt preserves evidence for the aggregation of some of the 5 terranes into superterranes or composite terranes, prior to the 1.1 billion-year-old orogeny. The early history of the Elzevir and Bancroft terranes shows many similarities to greenstone belt assembly in the Superior Province (Williams, Stott, Thurston et al., this volume).

- **Killarney Magmatic Belt**

The tectonic history of the Killarney Magmatic Belt (van Breemen and Davidson 1988) has been integrated into this tectonic overview of the Grenville Province.

- **Recognition of a variety of paleotectonic settings in the Central Metasedimentary Belt**

The review of the Grenville Province (*see* Easton, this volume) provides new insight into the variety of paleotectonic settings for Central Metasedimentary Belt metavolcanic and metasedimentary rocks, identifying oceanic plain, back-arc and platformal settings. Most Central Metasedimentary Belt metavolcanic rocks are tholeiitic, bimodal basalt-dacite sequences. True calc-alkalic rocks are uncommon in the Central Metasedimentary Belt.

- **Review of Grenville metallogeny within a modern tectonic framework**

An updated review of Grenville mineral resources, which includes industrial mineral deposits, is provided and deposit genesis is placed into a tectonic context. The mineral potential of the Central Gneiss Belt is evaluated for the first time.

- **Comparison of Grenville and Superior Province metallogenic patterns**

Similarities are seen in the paleoenvironmental setting of Grenville Province volcanic-associated base metal sulphide deposits and similar Superior Province deposits. Epigenetic gold deposits of both the Grenville and Superior provinces show similar relationships to regional structures and timing of mineralization.

- **Spatial and temporal patterns within the Central Gneiss Belt**

Patterns of spatial and temporal variation of plutonism within the Central Gneiss Belt are suggested. Some similarities to other Precambrian orogens are advanced.

- **Flinton Group–Grenville Supergroup relations**

The review of the Grenville Province concludes that the Flinton Group is indeed unconformable on the Grenville Supergroup and is a key unit in unravelling the history of the Central Metasedimentary Belt.

- **Orogenic metasomatic event**

The review of the Grenville Province (*see* Easton, this volume) confirms suggestions by Lumbers et al. (1990) and Lentz (1991) of the existence of a post-orogenic alkali-metasomatic event in the Bancroft–Pembroke area. This event generated many of the uranium, molybdenum and granite pegmatite occurrences in this region.

THE PHANEROZOIC EON: TECTONIC BASINS

Paleozoic and Mesozoic

The Paleozoic and Mesozoic eras in Ontario are represented by depositional sequences in sedimentary basins (*see* Figure 1.6), which formed during episodic inundation of Laurentia (*see* Figure 1.5) by inland seas. Episodic uplift and subsidence controlled relative sea level as well as rates of sedimentation of the clastic sedimentary rocks derived from the craton-margin orogenic belts and erosion of older cratonic rocks.

During the Paleozoic and Mesozoic eras, sediments ranging from at least the Early(?) Cambrian through to the Late Cretaceous periods were deposited in 4 major basins (Appalachian, Michigan, Hudson Bay and Moose River; *see* Figure 1.6), an intracratonic basin (Moose River Basin; *see* Figure 1.6) and in intervening platformal areas (Bally et al. 1989). Intermittent episodes of uplift and erosion during and subsequent to the Paleozoic and Mesozoic have resulted in reduced volumes, and a more restricted distribution of these sedimentary basin sequences, which presently cover approximately 220 000 km^2 in Ontario. Mainly basin margin sequences are represented in Ontario. Although thinner, and preserving a smaller interval relative to the basin centres, a maximum of 1525 m as compared to 4800 and 13 000 m in the centres of the Michigan and Appalachian basins, respectively, much of the stratigraphy is well exposed along eroded basin edges (e.g., the Niagara Escarpment, Figure 1.2).

The Paleozoic and Mesozoic geology of the province is described in a single comprehensive chapter (*see* Johnson et al., this volume) using a sequence stratigraphy approach. "Sequences" are unconformity-bounded depositional sequences of regional extent, more fully detailed in Johnson et al. (this volume). The Paleozoic and Mesozoic succession in Ontario is herein divided into 10 depositional sequences, that is, packages of genetically related sediments which are generally bounded by an unconformity. Some sequence boundaries are on a continent-wide scale similar to those of Sloss (1963, 1988), whereas other sequence boundaries reflect more local scale controls. Through this approach, a more detailed understanding of tectonic controls on sedimentation has evolved, with fundamental implications on the exploration and development of resources within these strata.

HIGHLIGHTS OF THE PALEOZOIC AND MESOZOIC OF ONTARIO

The major features of the Paleozoic and Mesozoic geology in Ontario are discussed in the chapter entitled Phanerozoic Geology of Ontario (*see* Johnson et al., this volume). Briefly listed below are the major advances in our understanding of the geological events which took place during these eras.

- **Comprehensive synthesis of Paleozoic and Mesozoic history of Ontario**

As the first province-wide synthesis of these eras, the integration of sequence stratigraphy and basin evolution permits identification of many new relationships, for instance, correlation of depositional sequences between the basins of southern and northern Ontario.

- **Sequence stratigraphy applied to Ontario Paleozoic/Mesozoic basins**

Ten major, largely unconformity bounded sequences have been described making up the record of Paleozoic/Mesozoic sedimentation in Ontario. Where depositional gaps exist, inferences have been drawn using strata of similar age in adjacent areas of New York and Michigan. This sequence stratigraphic approach has permitted recognition of the influence of distant orogenic events upon the types and distribution of sedimentary rocks in Ontario basins.

- **Units subdivided by tectonic affiliation**

The Paleozoic and Mesozoic sedimentary rocks have been classified by tectonic parameters such as craton-derived clastic rocks, orogen-derived clastic rocks, carbonates and evaporites. This subdivision has promoted identification of coherent large-scale tectonic "packages" of sedimentary rocks.

- **Sequence distribution maps**

Present and past distribution of major sediment types (orogen-derived clastic rocks, craton-derived clastic rocks, carbonates and evaporites) constituting each depositional sequence (as a "spatial slice in time") are presented. These maps represent the first comprehensive synthesis combining both northern and southern depositional basins. Depositional patterns on these paleogeographic maps illustrate: 1) evolution of sedimentary basins and interbasin arches, and their controls on sedimentation; 2) influx of orogen-derived clastic rocks during episodes of intense plate margin tectonism (i.e., major orogenies).

- **Foreland basins versus intracratonic basins**

The sequence stratigraphic approach highlights the contrasts in rock types, depositional environment and resource potential of the Appalachian foreland basin versus the intracratonic basins (e.g., Michigan Basin).

- **Identification and interbasin correlation**

Within the constraints of chrono- and/or biostratigraphic uncertainties, identification and interbasin correlation of sequence boundaries—either unconformities or significant changes in sedimentation—are described. Their tectonic significance is laid out for use in a comprehensive summary statement.

- **Integrated tectonic scenario**

This chapter provides a unified context within which cratonic uplift (epeirogeny), basin subsidence, plate margin tectonism and relative and eustatic sea level can be related (i.e., constraints on basin evolution models, controls on sea level curve modelling).

The Quaternary Epoch: Multiple Glaciations

The Quaternary deposits of Ontario are the product of multiple glaciations during the last 1.8 million years of Earth history. These glaciations left a legacy of tills, glaciofluvial, glaciolacustrine and fluviatile deposits. Tills are the most widespread glacial deposit type, with 3 end member compositions: sandy tills derived from erosion of Precambrian rocks, silty tills derived from erosion of carbonate rocks and clayey tills derived from incorporation of fine-grained glaciolacustrine sediments. In southern Ontario, numerous till sheets and other deposits of Illinoian and Wisconsinan age occur. The Quaternary sedimentary record is used to reconstruct, in particular, the advance and retreat of the Laurentide Ice Sheet. Glacial deposits in Ontario are dominated by materials deposited beneath the glacier (e.g., tills and eskers), along the ice margin (glaciofluvial ice-contact deposits) and clays and outwash channels related to proglacial lakes.

The chapter entitled Quaternary Geology of Ontario is accompanied by maps displaying the Quaternary deposits of Ontario at a scale of 1:1 000 000 (*see* Maps 2553 to 2556, map case).

ADVANCES IN UNDERSTANDING THE QUATERNARY OF ONTARIO

- **First provincial synthesis of the Quaternary**

The Quaternary chapter (*see* Barnett, this volume) and accompanying maps represent the first provincial synthesis of Quaternary geology for the whole province. Previous syntheses have concentrated on particular regions of the province (Chapman and Putnam 1984; Sado and Carswell 1987) or have been national in scope (e.g., Dredge and Cowan 1989).

- **Review of major landforms and glacial deposits**

The chapter provides a review of the genesis and character of the major types of glacially derived landforms and deposits, enabling the use of the chapter by nonspecialists. This approach will bring about a greater consciousness of Quaternary features on the part of mineral explorationists and other nonspecialists in Quaternary geology.

- **Southern Ontario Quaternary stratigraphy integrated with Quaternary history**

The chapter describes the major stratigraphic units integrated with the oscillations of the ice sheet and development of proglacial lakes.

- **History of proglacial lakes**

The history of the various proglacial lakes throughout the province is described and illustrated with a series of figures. This history allows the sequence of events forming major physiographic features throughout the province to be clearly developed. This history will be of benefit in education, exploration and land use planning.

- **Soil types and Quaternary history**

A brief section on the relationship between soil types and Quaternary processes and deposits allows the nonspecialist to understand some of the inter-relationships between these related fields.

- **Quaternary geology and resource extraction**

Quaternary deposits are an unappreciated source of a large proportion of the wealth derived from Ontario's geological heritage. The chapter provides an overview of, and specific examples of the types of Quaternary units from which aggregate deposits are extracted. Features controlling the economics of aggregate extraction are also described.

- **Quaternary deposits and mineral exploration**

The principles of Quaternary geology in mineral exploration are described in a brief section which will prove invaluable as an introduction to important literature for those geologists coping with exploration through a veneer of Quaternary deposits.

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A project of this type has never been previously attempted by the Ontario Geological Survey. Therefore, even though we attempted to lay the course out before starting the work, a lot was learned by actually doing the work. Therefore, the various authors and indeed all staff associated with the project exercized much patience, while the editors attempted to find their way through the many problems endemic to large projects.

The project required a high level of technical support through its entire course. The individuals who carried out the assembly of background information, technical editing of draft manuscripts within the operating sections of the OGS, and the assembly of map manuscripts by many authors are "behind the scenes" workers who made an enormous contribution to the success of the project. We acknowledge, therefore, the hard work of Kathy Chivers-MacMillan, Susan Cole, Michele Dolan, Angela Gulley, Sean MacGowan, Jeanine McRoberts, Sandra Neufeld and Ikram Osmani. Design of the publications and the monitoring of the publication process required the Publication and Cartographic Services Unit to solve many new problems; especially invaluable were Ray Balgalvis, Guy Kendrick and Tammy Hoey. Richard Cherer kept our computer systems up and running despite our best efforts to overload them during the writing and editing stages.

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APPENDIX

These definitions are adopted for consistency within the *Geology of Ontario* volume. Past and future usage within the OGS and scientific community may differ. Informal "phraseology" is adopted for clarity.

Craton

A craton is defined by the AGI Glossary (Bates and Jackson 1987) as: " A part of the Earth's crust that has attained stability, and has been little deformed for a prolonged period."

Basin

A sedimentary basin is an area underlain by a substantial thickness of sedimentary rocks. Thickness is generally in the order of several kilometres.

Regolith

The term regolith is a general term (Bates and Jackson 1987) for "the layer or mantle of fragmental and unconsolidated rock material whether residual or transported and of highly variable character that nearly everywhere forms the surface of the land and overlies or covers the bedrock. It includes rock debris of all kinds, volcanic ash, glacial drift, alluvium, loess and eolian deposits, vegetal accumulations and soil."

Shear Zone

A shear zone is a zone of bedrock, with subparallel walls, in which high strain is localized. In the absence of fracture or fault planes, such a zone is known as a ductile shear zone. Where the zone contains discrete fracture and/or fault planes, it is known as a brittle-ductile shear zone. Most observed shear zones in the Superior Province of Ontario display brittle-ductile to ductile features. Faults are brittle shear zones across which displacements are recorded (*modified from* Ramsay 1980; Ramsay and Huber 1987). Discrete shear zones or brittle faults are generally less than 10s of metres in thickness and may be kilometres to 10s of kilometres in strike length.

Deformation Zone

A deformation zone is a broad, generally curviplanar zone of high strain. It consists of numerous discrete, narrower zones of high strain, which are separated by domains that exhibit little deformation. An envelope, drawn to enclose the highest density of discrete high-strain zones, constitutes the deformation zone boundaries. Deformation zones are generally 100s of metres to kilometres in thickness, and can extend for 10s to 100s of kilometres along strike length (*modified from* Colvine et al. 1988; Hodgson 1990).

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Chapter 2

The First Hundred Years: A Brief History of the Ontario Geological Survey

E.G. Pye

Former Director, Ontario Geological Survey

Abstract

The Ontario Geological Survey, from its establishment in 1891, has made considerable contributions to the mineral industry and economy of the province of Ontario. Early field investigations not only yielded the first geological maps of the province, but set the pace for the high standard of research that is associated with the Survey today.

In the early years, the Survey had limited staff and regularly employed university personnel to carry out field studies. The Survey was involved from the outset in service to prospectors and the burgeoning mineral industry. This early pattern of service to the exploration industry remains a principal objective.

During the 1930s, the increase in the price of gold signalled the beginning of the Survey's concentrated investigation of Archean greenstone belts in northern Ontario. However, with the outbreak of WW II and the introduction of the Canada–USA lend-lease program, priorities switched to base metals and strategic materials, especially in southern Ontario. It was the search for magnetic ores that, in 1948, prompted the Survey to carry out its first airborne magnetic survey, eventually leading to federal–provincial agreements to map the entire province in the 1950s and 1960s.

In the early 1940s, the first Resident Geologist's office was established under the Survey's direction in Thunder Bay; and in the early 1950s, a comprehensive study of the province's industrial mineral resources was initiated.

The early 1960s witnessed an increase in the number of staff employed by the Survey; and successive reorganizations in the 1960s and 1970s resulted in formation of the Precambrian Geology, Engineering and Terrain Geology, Mineral Deposits Geophysics/Geochemistry and Geoservices (support services) sections. In 1978, the Mineral Services (Laboratories) Branch of the Ministry of Natural Resources was incorporated into the organization, which was now formally recognized as the "Ontario Geological Survey" by the Ontario government.

In the 1970s, the Survey initiated the Ontario Geoscience Research Grants program to encourage applied research at Ontario universities, and a complementary Exploration Technology Program to assist the private sector in developing new exploration equipment and field methods. The Survey also took steps to establish a world-class geochronological facility, the Jack Satterly Geochronology Laboratory, at the Royal Ontario Museum. In the late 1970s and early 1980s the Survey initiated a drill core storage program, perhaps the most comprehensive such program in North America, with warehouses at Bancroft, Kenora, Sault Ste. Marie, Swastika, Thunder Bay, Timmins and Tweed.

The Ontario Geological Survey also carried out a number of specialized research projects during the 1970s and 1980s: a study of the stratigraphy and limestone/dolostone resources of Manitoulin Island; engineering geology terrain studies in northern and southern Ontario; several multidiscipline community-based projects such as the Kirkland Lake Initiatives Program and a number of programs funded jointly by the federal and provincial governments; a Hydrocarbon Energy Resources Program; and a detailed government-industry-university study of the Sudbury Structure as an Ontario Bicentennial project. An important result of the work in northern Ontario was an increased understanding of the Archean achieved through detailed studies in the Superior Province by staff geologists.

Today, advanced exploration techniques such as geochronological dating and geochemical and geophysical surveys, have placed the Survey in the forefront of research on the historical, structural, and economic geology of the province.

The Ontario Geological Survey, as in the past, will continue to work closely with industry and the academic community to enhance the future economic and environmental well-being of Ontario. The Survey also continues to offer significant employment and career opportunities for Canadian geoscientists.

INTRODUCTION

The Ontario Geological Survey formally came into being in 1978 with the amalgamation of the former Geological Branch and Mineral Research (Laboratory) Branch of the Ontario Ministry of Natural Resources (Pye 1985). The Ontario Geological Survey dates its origins, however, to the recommendations made by a Royal Commission on the Mineral Resources of Ontario (the Charlton Commission) appointed by the provincial legislature in 1888 (Report of

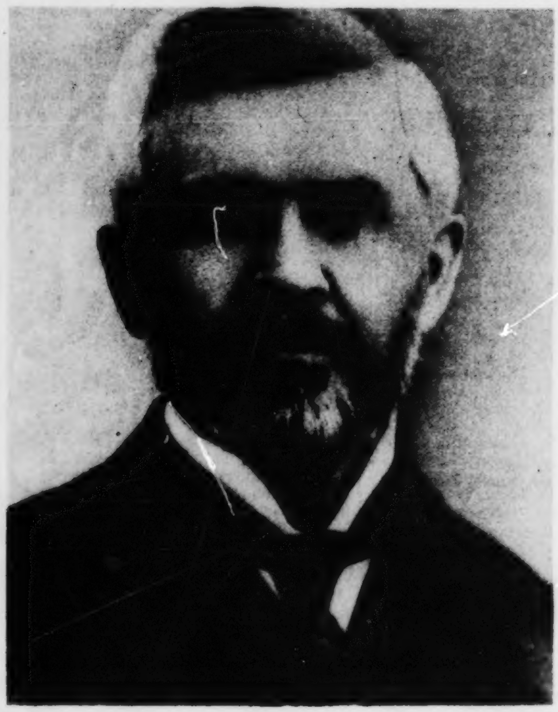

Figure 2.1. Archibald Blue, Director of the Ontario Bureau of Mines from its foundation in 1891 to 1900 (Gibson 1933).

the Royal Commission on the mineral resources of Ontario and measures for their development 1890).

The legislature, "shocked by the revelation of the appalling suffering endured by women and children in the British coal mines" (Smith 1986, p.90), and concerned with the health and safety of Ontario mine workers, had been strongly influenced by a spate of mineral discoveries made in the province in the 1870s and 1880s. Gold had been discovered on the Richardson farm near Madoc in 1868. With the California and British Columbia gold rushes still in recent memory, this discovery set eastern Ontario "on fire" (Gibson 1937, p.4). Additional discoveries (including the Cordova and Delora mines in eastern Ontario) added fuel to the "fire". In the early 1870s, Peter McKellar located the Huronian (Moss) gold mine southwest of Kashabowie in the Rainy River District of northwestern Ontario; his brother Archibald McKellar discovered a gold-bearing quartz vein west of Lac des Milles Lacs, and another brother, Donald McKellar, found a gold deposit at Jackfish Bay, on Lake Superior (Gibson 1937, p.5). In the late 1870s, gold also was discovered in the Lake of the Woods and Sudbury areas (Gibson 1937, p.5). Later, in the 1880s, nickel-copper deposits were found in the Sudbury area, commencing with the great Murray Mine in 1883 (Gibson 1937, p.78).

In keeping with the Commission's report of 1890, the Ontario legislature created the Ontario Bureau of Mines on May 4, 1891 as a division of the Department of Crown Lands to "aid in promoting the mining interests of the Province." Archibald Blue (Figure 2.1), a self-educated civil servant and one keenly interested in Ontario's minerals industry, was appointed the Bureau's first director.

THE FIRST DECADE

Archibald Blue, and his successor in 1900, Thomas W. Gibson (Figure 2.2), following the Charlton Commission's recommendations, put major emphasis on geological investigations. Without any geologists on the staff at the time, one can only speculate that either complement positions were not provided or that suitable people were not available, the Bureau was fortunate in being able to employ, under contract, university personnel, for example, A.P. Coleman and W.A. Parks of the University of Toronto and W.G. Miller of Queen's University, during the summer months. This practice of employing university personnel during the summer months was successfully continued until the early 1960s. A.P. Coleman's first concern, on behalf of the Bureau in the 1890s, was in relation to iron ores. He described Ontario's iron ore samples displayed by the Bureau at the Chicago World's Fair (Coleman 1893a, p.78–82), separately from other minerals which were also on display (Coleman 1893b, p.185–194) to promote interest in Ontario's fledgling mining industry. In 1894, Coleman reported on building stones,

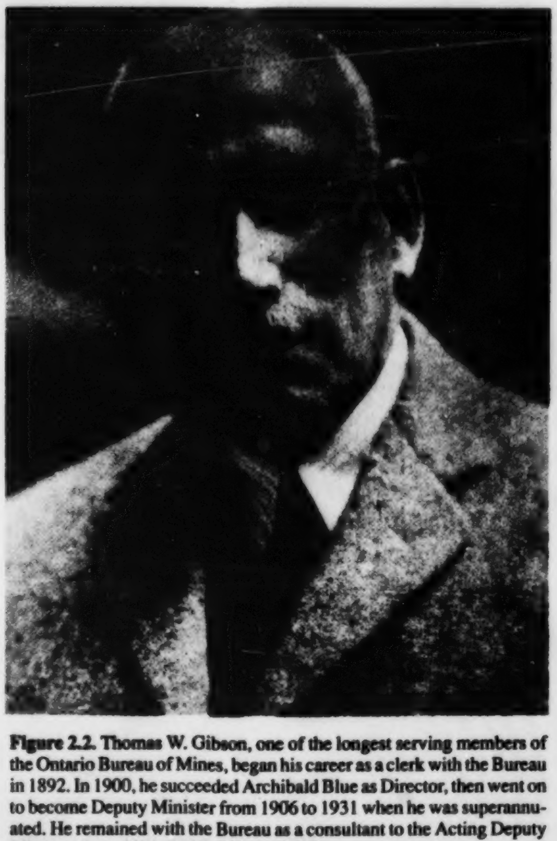

Figure 2.2. Thomas W. Gibson, one of the longest serving members of the Ontario Bureau of Mines, began his career as a clerk with the Bureau in 1892. In 1900, he succeeded Archibald Blue as Director, then went on to become Deputy Minister from 1906 to 1931 when he was superannuated. He remained with the Bureau as a consultant to the Acting Deputy Minister from 1931–1935 (Gibson 1933).

and in 1895–1896 he investigated the gold deposits of the Rainy River District in northwestern Ontario. However in 1898, he returned to his interest in iron ores, and from then until the end of the decade studied iron ranges in northwestern Ontario, in particular, the Michipicoten range at Wawa Lake (Coleman 1899, p.121–174).

W.A. Parks (1898, p.161–183), a colleague of Coleman at the University of Toronto, was employed by the Bureau in 1897 to carry out geological reconnaissance surveys along base and meridian lines in the Rainy River District in support of Coleman's work. Later, following Coleman to the Michipicoten area, Parks (1899, p.175–176) carried out a reconnaissance survey, started by E.M. Burwash in 1896 (Burwash 1897, p.167–184), along the Nipissing–Algoma meridian line.

While Coleman and Parks were busy in northern Ontario, W.G. Miller was engaged from 1898 to 1899 to study the wide diversity of industrial minerals, which since the gold discovery on the Richardson farm in 1866, had been found in Ontario.

In its report, spurred by W.H. Merritt, later of the Kingston School of Mining affiliated with Queen's University (Figure 2.3), the Charlton Commission concluded:

> In order that the mineral resources of the province may be successfully and economically developed it is desirable that measures should be taken for the practical and scientific training of all who may engage in the industry. Prospectors and explorers are found to be very deficient in the kind of information which would enable them to prosecute their arduous labours to best advantage; and your Commissioners recommend for that purpose the adoption of a scheme [short classes for miners] such as has been tried with gratifying results in the colony of New Zealand.

A pioneer prospector's class was held by W.H. Merritt at Marmora in 1894 with such success that prospectors' (Mineral Exploration) classes have become a tradition of the Survey to the present day (Pye 1989) (Figure 2.4).

THE "ERA" OF WILLET GREEN MILLER

In its report of 1890, the Royal Commission on the Mineral Resources of Ontario had emphasized, on the grounds that the only information available on the geology of the province resided with the Geological Survey of Canada, that the time had come for the province to "undertake a...detailed and utilitarian survey of its own [because] there are numerous districts of great economic importance in which little or no detailed work has been done, and in regard to which very little systematic information is available." The Commission also recommended that this survey be directed by a geological officer who "should be well versed in the geology of the province and should have a sufficient knowledge of mines and mineral deposits in general to enable him to give useful advice, when desired, to those actually engaged in mining, or to persons proposing to work undeveloped properties. From his responsible office and neutral position such advice would naturally carry more weight than that of experts employed as special leaders, no matter how well qualified they may be."

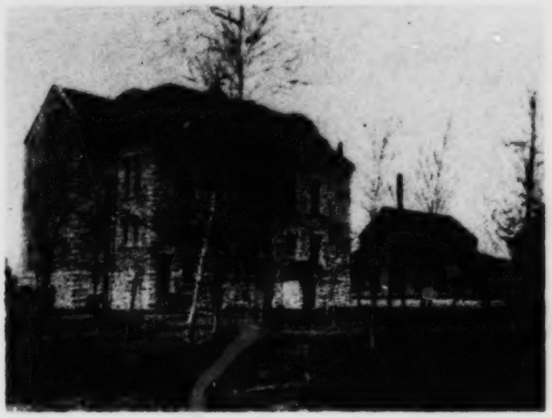

Figure 2.3. The Kingston School of Mining, one of Ontario's first institutions for training geologists and prospectors; consisted of the John Caruthers Science Hall (left) and the Mining Laboratory (right). The only other geological institute, at this time, was the School of Practical Science in Toronto (anonymous photograph 1896).

Figure 2.4. A prospectors' class in Sault Ste. Marie during the 1970s. E.B. Freeman, of the Ontario Geological Survey (left), exchanges information and ideas with interested onlookers (photograph *courtesy of* the Ontario Geological Survey, Mines Library).

In 1902, Willet Miller (Figure 2.5) of Queen's University, was appointed as Ontario's first Provincial Geologist together with 3 assistants, Cyril W. Knight, A.G. Burrows and Percy E. Hopkins (Gibson 1937, p.11). Despite an increased annual budget, the practice introduced by

Figure 2.5. Willet Green Miller (standing) was the Ontario Bureau of Mines' first provincial geologist from 1902 to 1925. The identity of the seated man is unknown (photograph *courtesy of* the Ontario Geological Survey Mines Library).

Archibald Blue of hiring university personnel in the summer months, fortunately was maintained, and the work of Miller and his assistants continued to be supplemented by that of Coleman, Parks and others.

The late J.E. Thomson, director of the Ontario government's Geological Branch from 1966 to 1972, presented an excellent summary of Miller's career, within both the Survey itself and the geological community at large (Thomson 1970a, p.5). Miller is best remembered as "the father of Cobalt". According to Thomson, his intimate connection with the discoveries of silver there tended to become obscured by legends, fictions and fantasies: "He first visited the area in 1903 to check on the source of a sample of niccolite that had been presented to T.W. Gibson. He was shown four veins discovered along the new Timiskaming and Northern Ontario Railway. Three of the veins were found to be 'very rich in native silver'." According to Thomson (Thomson 1970a, p.5), "Chunks of native silver as large as stove-lids or cannon balls lay on the ground around the showings but the blackened, tarnished ore had attracted little or no attention...[Miller] soon gave the news to the world through an article published in the *Engineering and Mining Journal* and a report of the Ontario Bureau of Mines". The rush which followed, the growth of the Cobalt camp and expansion of Ontario's mining industry from that point on are well known (Smith 1986, p.133–151; Hanula and Longo 1982, p.184–189).

Over the next several years, Miller and his assistants continued to study the silver deposits and map the Cobalt area. So much in demand was Miller's report on the "*Cobalt-Nickel Arsenides and Silver Deposits of Timiskaming, Ontario*", first released in 1905, and so quickly was new information obtained, that a second edition was published in 1907, a third edition in 1908 and a fourth edition in 1910.

Another of Miller's more important contributions to the mining industry of the province was early recognition of Cobalt as a "shallow" camp. According to Thomson (Thomson 1970a, p.5):

> He soon recognized that most of the silver ore would be found in the veins occurring in the brittle conglomerate but these would soon die out when they entered the tougher underlying greenstone. These early predictions...were abundantly borne out by later mining operations. So, a large part of the mining profits that might have been wasted on deep exploration at Cobalt were instead diverted to the development of other new, neighbouring camps, such as Porcupine.

On Miller's appointment in 1902, A.P. Coleman was temporarily diverted from his studies of Ontario's iron ranges to the Sudbury area, which was becoming increasingly important to the provincial economy. No geological map of the Sudbury area existed in the 1800s at the time of the first discoveries (Giblin 1984, p.6–7). The area was first systematically mapped for the Geological Survey of Canada by R. Bell (1891), who reported that the nickel-copper deposits were associated with 2 ranges of mafic rocks on either side of an easterly trending geological basin. Coleman, on the other hand, showed conclusively that the mafic rocks, with which the ores were associated, extended continuously around the basin as a "laccolithic sheet". Following 3 summer's work in the area, Coleman (1905, p.2–19), concluded that the nickel-copper deposits were of magmatic origin and occurred at the basal margin of the intrusion or in offset dikes radiating outward from the intrusion. These findings were to prove of inestimable value in directing exploration in the area for many years, and were among the Survey's most important contributions to the mining industry of the province throughout the past century.

After completing his final report on the Sudbury area, Coleman returned to his investigations of Ontario's iron ranges. He continued his work in the Michipicoten area in 1905, and together with E.S. Moore, his protégé at the University of Toronto, then began mapping the iron ranges east of Lake Nipigon (Coleman 1907; Moore 1907).

In 1912, Miller was faced with a new challenge, for in that year what is now the Patricia Portion of the District of Kenora was added to the territory of the province by an Act of the Federal Parliament. Despite his small staff and meagre budget, Miller met this challenge by arranging with the Geological Survey of Canada to republish a number of the latter's reports of reconnaissance surveys made previously by Dowling, Bell, Camsell and others. In the Introduction to the Ontario Bureau of Mines Annual Report for 1912 (Miller 1912), Miller stated:

> From time to time during the last forty years the Geological Survey of Canada has sent parties to explore and report on that part of the district of Keewatin to which the name Patricia has been

> given. Much valuable information has been obtained by these parties, but, being scattered through the annual reports of the Geological Survey, many of which are out of print and difficult to access, it is not available to the public. Hence it was decided to collect and reprint the reports, with certain of their accompanying maps and illustrations, and to publish them in a single volume, so that officials of the Ontario government, prospectors for minerals, and other persons visiting the district, can have in handy form practically all the information that has been obtained.

A.G. Burrows commenced mapping in the Porcupine area in 1911 (Burrows 1911) following the new gold discoveries of that year, starting a program of mapping of greenstone belts with potential for gold mineralization by provincial government geologists extending over the next 50 years. With war clouds gathering in Europe, however, the emphasis Miller had placed on Cobalt, Sudbury, the iron ranges, and the new gold camps, soon shifted to Miller's "old stamping ground" in eastern Ontario, where there were possibilities of finding materials needed for the war effort. This change was short-lived, and by the early 1920s, most of the work of the Survey again was directed to elucidating the geology of known and potential gold camps (Figure 2.6).

A feature of the Miller "era" is the wide diversity of interests demonstrated by the Survey's publications. As one might expect, most reports deal with mineral resources of direct economic interest in strict keeping with the Bureau's mandate to "aid in promoting the mineral interests of the Province". However, Quaternary geology (Coleman 1904, 1909, 1913; Taylor 1913) and paleontology were not neglected (Whiteaves 1912; Wilson and Mather 1916; Parks and Dyer 1921), and even botany (Macoun 1912) found its way into the Bureau's reports. It was an auspicious beginning.

Miller died in 1925. He was succeeded as Provincial Geologist first by A.G. Burrows, later in 1934 by H.C. Rickaby; in 1938, by M.E. Hurst; in 1966, by J.E. Thomson; in 1972, by E.G. Pye; and in 1984, by V.G. Milne.

THE INTER-WAR YEARS

Immediately before and during World War I, a large part of the mineral production of the Sudbury and Michipicoten areas went into armaments. When the war ended, however, the nickel-copper mines at Sudbury found themselves with a much smaller market, surplus productive capacity, large stocks in inventory, and an expanded and largely unneeded work force. The market for nickel had dwindled from 92 million pounds in 1918 to only 6.5 million pounds in 1921 (Smith 1986, p.213). International Nickel was forced to close its Sudbury mines and its Port Colborne and New Jersey refineries. The Helen iron mine at Wawa had already closed. The only bright "spot" for the industry was gold mining, reflecting a well-established and stable price of $20.67 per troy ounce. Exploration soon resulted in several important discoveries as prospecting intensified: the Howey and

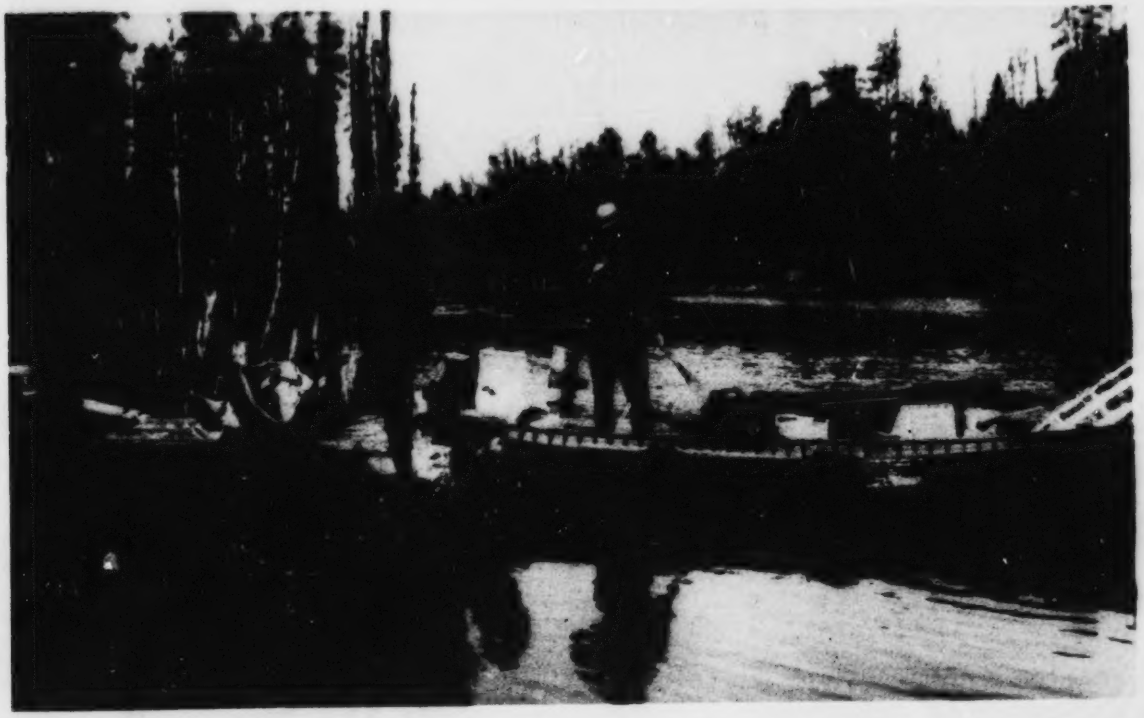

Figure 2.6. A geological field party with their gear stop at Borland's Half-way House on the way to Shining Tree, northeast of Wawa. This area was investigated for its gold potential during the early 1920s (photograph *courtesy of* the Ontario Geological Survey Mines Library).

Cochenour–Willans discoveries in the Red Lake area in 1926 (Smith 1986, p.226–227); the Pickle–Crow and Central Patricia discoveries in the Crow River area in 1928 (Gibson 1937, p.27–28); the Hard Rock and Little Long Lac discoveries in the Little Long Lac area in 1931 (Gibson 1937, p.33–34); and the Bankfield discovery in the same area in 1932 (Gibson 1937, p.35).

With the onset of the Great Depression of the thirties, President Roosevelt, in attempting to get the stagnant economy of the United States moving, raised the price of gold to $35.00 per ounce. The new price and the upsurge in exploration "brought about an enormous expansion of the industry. From 1934–1940, 40 new mines were opened and a dozen old ones were revived" (Smith 1986, p.255). In his 1937 report Gibson (1937, p.36) said: "The mining scene in Ontario is widening at a rate beyond the ability of any chronicler to keep pace with it. Statistics and statements become obsolete before they can be put into print. New gold fields come into view year by year, new mines are opened, and production rises rapidly."

The decline in the importance of base metals and the rise in the importance of gold to the economy in the 1920s and 1930s had a corresponding effect on the geological mapping program of the Ontario government. This program, while it did not neglect studies of the Sudbury area, the silver fields and the iron ranges, at the same time tended to shift away from these areas and concentrate on the Archean greenstone belts where the new gold discoveries were being made (Figure 2.7). One recalls the reports and maps of Burrows, Coleman, Hopkins, Knight, Miller, Moore, Parks, Parsons, Rickaby and Todd of the early part of the period; of Baker, Bruce, Gledhill, Hurst and Phemister of later years; and more recently of Harding, Horwood, Moorhouse, Satterly and Thomson. Of the many geological reports of the period, particularly impressive were those of Todd (1928) on the Kirkland Lake area and Horwood (1940) on the Red Lake area. Also noteworthy was Thomson's (1932) report on the Heron Bay–White Lake area and accompanying map which identified the sulphide deposit that in the early 1950s was staked to become the Geco Mine at Manitouwadge (Pye 1957, p.3–4).

Field Work in the 1920s and 1930s

In the early days of the geological mapping program in Ontario, good topographic maps were wanting. In the more settled areas of the province, ground control for geological mapping was provided by township, lot and concession lines. However, in unsettled areas, particularly in northern Ontario, survey lines were sparse. Here mapping was done along the few base and meridian lines, and along the principal canoe routes. Projects were as much topographic surveys as geological, the geologists preparing their own base maps as they travelled (Figure 2.8).

In the interwar years, with the advent of aerial photography and improved base maps, pace-and-compass traversing between topographic features recognizable on airphotos came more and more into vogue. For detailed surveys in the mining camps, mapping also was commonly done along claim lines by using compass or plane table and open-sight alidade between the lines.

Figure 2.7. Cyril W. Knight (front centre) with a geological and topographical survey party in camp on Misema Lake, near Kirkland Lake area in 1920. Because of the increasing importance of gold, many such field parties were sent to investigate potential gold areas in northern Ontario (Knight 1920).

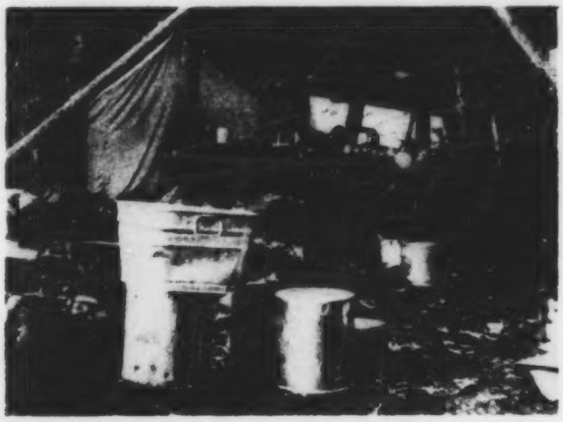

Figure 2.8. A typical survey party kitchen at the Algoma–Thunder Bay District boundary meridian line, 1949. A camp such as this would have been much the same during the early 1900s when survey lines were used as controls in geological mapping (photograph *courtesy of* the Ontario Geological Survey Mines Library).

Today, nearly all maps areas are accessible by aircraft or automobile. In the years prior to World War II, however, aircraft were not readily available in the north, and roads were few and rough by present standards (Figure 2.9a and 2.9b). Highways between the principal centres were either nonexistent or only partly completed. The Trans–Canada Highway was still nothing but a dream. As in the earlier period, the geologists generally had to travel by train (Figure 2.10) to a town or siding, sometimes getting the conductor of a way freight to stop at a river crossing, from which the field party would then travel by canoe to the map area, frequently carrying heavy loads over long portages. Fresh food was often a luxury, and a monotonous diet could only be relieved by fishing or by arduous trips back to "steel". As the reader is fully aware, this was soon to change. J.E. Thomson (1971, p.15) described his 1929 field season in the Pickle–Crow area as follows:

> We went in from Savant Lake station by a canoe route that took 12 days of hard going. At the end of the season we came out by plane in one hour [Figure 2.11]. This was the first time the [Ontario] Department [of Mines] had used aircraft for travelling although most of our supplies were flown in during the summer...We lived on very plain fare—mostly bannock, bacon, beans, rice and 'bully beef' with plenty of oatmeal porridge. Dr. Hurst [party leader] made the bread while I made the bannock. It was difficult to judge which was most indigestible.

Figure 2.9a. Two methods of travel in northern Ontario in the days before WW II. An early road to Shining Tree, northeastern Ontario.

One must not conclude that field work has become a "cinch". Geologists still often hike or paddle long distances before starting a traverse, and going through alder swamps and areas of hilly topography is no less difficult, and the flies and mosquitoes are no less pesky. But working and living conditions have improved well beyond the dreams of the 1920s and 1930s (Figure 2.12a and 2.12b). This observer still marvels at the general excellence of the maps produced during that period.

In the early years, the maps published by the Survey were primarily lithological maps, emphasizing the areal distribution of rock types rather than stratigraphy (Pettijohn 1984, p.244). The maps published by the Survey in the 1930s, however, were both unusual and innovative for the time, both geologically and cartographically. Unlike those of earlier days, they began to show symbols indicating the facing directions of volcanic and sedimentary rock units in the Archean greenstone belts and other features such as schistosity. The maps also began to represent rock formations in two-tone colours to distinguish outcrop areas from areas of inferred geology. According to Pettijohn (1984, p.247):

> Mapping techniques have...changed. The use of two-tone colours to discriminate between actual outcrops and inferred distribution of formations...became common in the Ontario Department of Mines about 1936. Only since the early sixties has the United States Geological Survey used this technique. I would regard large-scale maps of the Canadian and Baltic shields that failed to make this

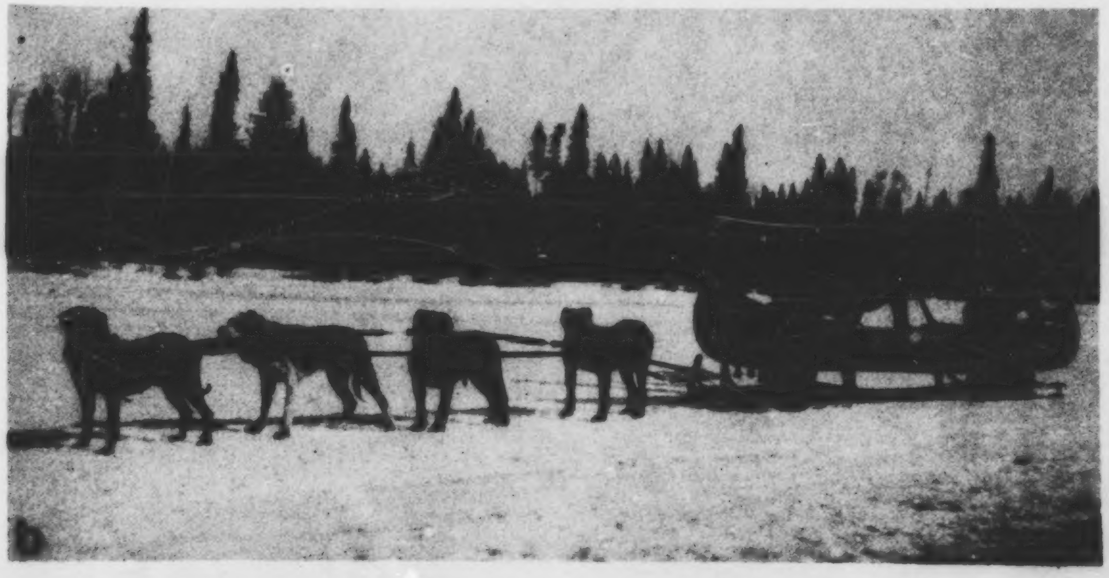

Figure 2.9b. Geologists with a dog sled on Porcupine Lake, northeastern Ontario. During winter months, a frozen lake or river provided an abundant alternative to roads which were few and far between (photographs *courtesy of* the Ontario Geological Survey Mines Library).

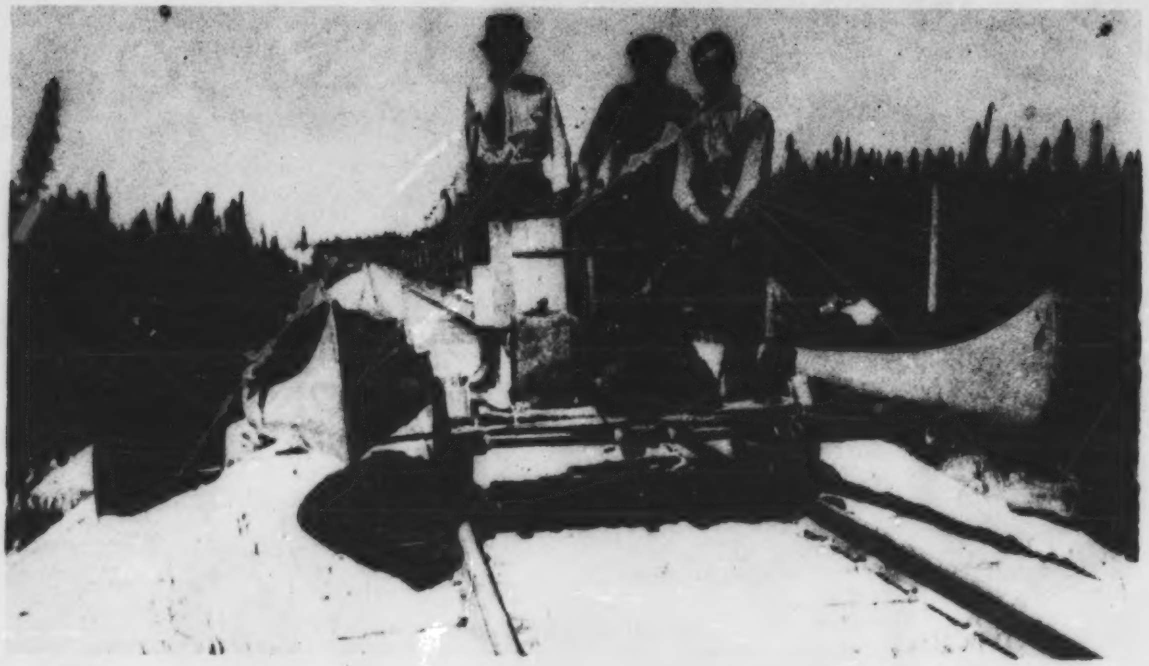

Figure 2.10. Travelling by rail line was often used in place of roads. However, access to the map area by rail was limited and canoes had to be relied upon for the remainder of the trip. Location and identity of persons in the photo are unknown (photograph *courtesy of* the Ontario Geological Survey Mines Library).

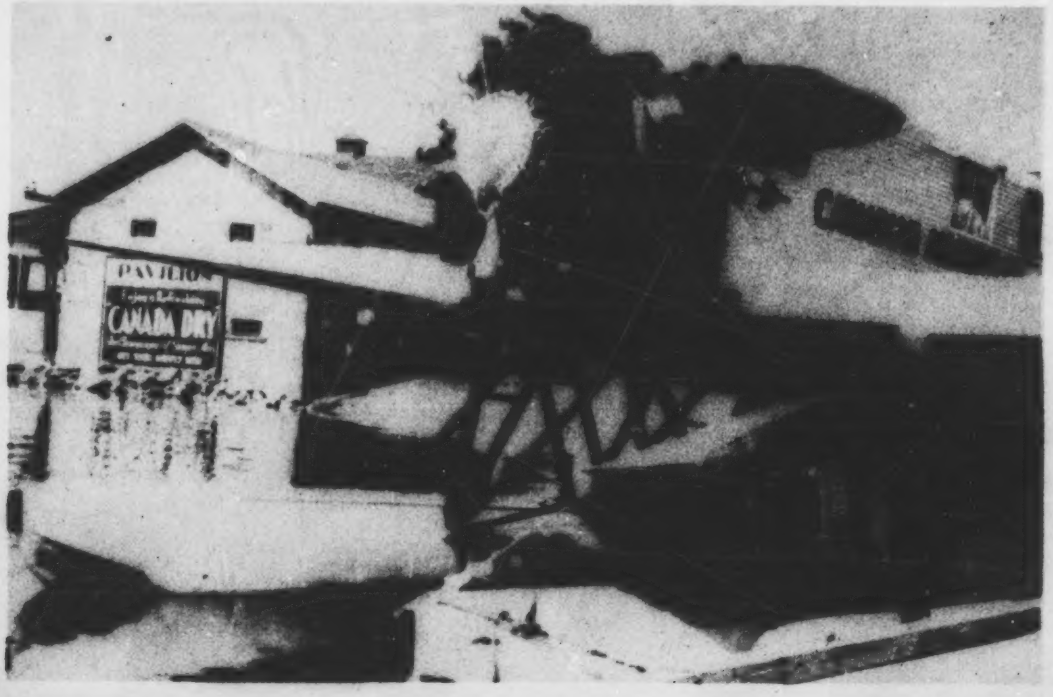

Figure 2.11. A Junkers ASN bush plane at Sioux Lookout, 1937, is ready to fly the field crew of Jack Satterly into a remote map area. A canoe strapped to the pontoons remains a common sight today (photograph *courtesy of* J. Satterly).

distinction as obsolete and unsatisfactory even if constructed today. The Canadians also made a significant advance when they showed, by appropriate symbols, the top position of sediments and lavas where observed. Such is not yet generally done. The practice of showing the criterion (graded or cross-bedding, etc.) and the place where seen by specific map symbols seems again to have been initiated by the geologists of the Ontario Department of Mines.

Figure 2.12a. Campsites of the 1930s. a) Early morning at a campsite on Caribou Lake, northern Ontario in 1938. Today, portable propane stoves and heaters have replaced the campfire for cooking and warmth. b) Base camp on Weagamow Lake, northern Ontario, 1938 (geological assistants from left to right: Mac Robson, Karl Bayly and John Moore). Although the government geologists of today have a greater variety of equipment, the canvas tent is still widely used as living quarters in the field (photographs *courtesy of* J. Satterly).

WORLD WAR II AND THE 1940s

Shortly after the outbreak of the war in Europe, at the 1940 convention of the Prospectors and Developers Association, Paul Leduc, Minister of Mines for Ontario, announced that the Department's geologists had been advised to give special attention to strategic war minerals, and that he considered gold to be "the prime war metal of all" (Hanula and Longo 1982, p.43), essential for the purchase of military hardware from the United States. Gold mining reached its peak in 1941, and thence declined rapidly owing to shortages of men and materials. It was particularly hard hit by the 1941 lend-lease agreement which rendered gold no longer important to the war effort. From this juncture, the priorities of the industry switched to base metals and strategic minerals (Hanula and Longo 1982, p.46–47).

Accordingly, the priorities of the Department also changed. As in World War I, emphasis again shifted to southern and eastern Ontario. Thomson (1943) reported on the north Hastings area; Satterly (1944) on the Renfrew area; Meen (1944) on the Grimsthorpe–Barrie area; and Harding (1944a, 1944b) on Kaladar and Kennebec townships and the Mattawan–Olrig areas. At the end of the war, however, the Survey's mapping pendulum swung back to northern Ontario.

The depression which was expected after the war, as happened after World War I, did not materialize. On the contrary, the country entered a period of postwar reconstruction and unprecedented industrial expansion. While gold mining remained in the doldrums and depended upon the federal government's program of Emergency Gold Mining

Figure 2.12b. Base camp on Weagamow Lake, northern Ontario, 1938 (geological assistants from left to right: Mac Robson, Karl Bayly and John Moore). Although the government geologists of today have a greater variety of equipment, the canvas tent is still widely used as living quarters in the field (photographs *courtesy of* J. Satterly).

Assistance for its continued existence, demand for base metals continued to grow. Moore and Armstrong (1946) made a detailed survey of the iron range in the Michipicoten area for the Department. The Department of Mines, however, did not otherwise respond immediately to the new reality at war's end; and mapping of the favourable gold areas, started before the war, was resumed. Thomson (1950) returned to his detailed mapping of the Kirkland Lake area; Horwood (Horwood 1940; Horwood and Pye 1951) commenced work in the Little Long Lac gold area; Satterly (1949) began mapping in the Cochrane District along the Porcupine–Destor "break"; and Hewitt (1951) mapped Skead Township in the Larder Lake area.

An important scientific development during World War II was the invention of the airborne magnetometer to detect enemy submarines. The Department was quick to realize the significance of this to the prospecting for magnetic ores. In 1948, on the initiative of the Harold C. Rickaby, Deputy Minister, the Department's first airborne magnetic survey was made of Hastings County in eastern Ontario, where magnetic iron deposits had been discovered in the early 1800s. The survey identified a large "thumb-print" anomaly at Marmora. This anomaly was quite similar to one produced by a new discovery in Pennsylvania, and attracted the attention of the Bethlehem Mines Corporation. A drilling program by Bethlehem Mines Corporation in 1950 outlined an ore body of 20 million imperial tons, which was brought into production in 1955 (Smith 1986, p.307). The Hastings County survey was the start of many such surveys in co-operation with the Geological Survey of Canada (*see* section entitled "Airborne Magnetometer Surveys").

In 1944, at its annual meeting, the Prospectors and Developers Association recommended that Resident Geologists be established in the country's principal mining camps (Hanula and Longo 1982, p.50–51) to collect and disseminate information on local mineral deposits and exploration activities, and provide advice to prospectors. The first such office was established by the late H.C. Horwood in the District Court House in Port Arthur (now Thunder Bay), as being strategically located with reference to the gold camps of northwestern Ontario. Later, additional offices were established in Kenora (1944), Kirkland Lake (1945), Timmins (1946) and Cobalt (1947). Also at this time, the field survey staff at head office in Toronto was increased from the original 3 allocated to Miller in 1902 to 5. In today's context, this seems little, but with the Great Depression still in recent memory, and the success that the Department of Mines had in employing highly qualified university personnel to assist with its mapping, this addition was generous indeed. In the 1930s, the 3 positions first held by Burrows, Hopkins and Knight were filled by Harding, Satterly (Figure 2.13) and Thomson. In the late 1940s, Horwood was transferred to Toronto from Port Arthur, (his position as Resident Geologist being taken by M.W. Bartley).

Figure 2.13. Jack Satterly with several field crew members (left to right: Mac Robson, Cam Robertson and Jack Satterly). Jack Satterly later became one of the Department of Mines' 5 senior geologists during the 1940s. In 1977, the Jack Satterly Geochronology Laboratory at the Royal Ontario Museum, was named in his honour (photograph *courtesy of* J. Satterly).

THE EXCITING 1950s

Metallic Minerals

The 1950s were exciting and fruitful for Ontario's mining industry, as one important discovery followed another. The Ontario Department of Mines, with its limited resources, had difficulty keeping pace with the demand for new information required by the prospector. With the discovery of the Tribag copper deposit, Thomson (1954) carried out a detailed survey of the Mamainse Point area north of Sault Ste. Marie on the east side of Lake Superior; and with the growing importance of low-grade "concentrating" iron ores, and growing interest in a deposit at Whitefish Lake southwest of Port Arthur, Moorhouse (1960) was engaged to map the Gunflint iron range. Early in the decade, Satterly, joined by V.K. Prest and others, continued detailed mapping in the Cochrane District north of Kirkland Lake.

The reader will be aware that World War II ushered in the atomic age. In the war years, because of the "Manhattan project" that led to the atomic bomb, prospecting for uranium in Canada was restricted to the Crown Corporation, Eldorado Mining and Refining Limited. In 1947, however, with the demand for uranium increasing, and no new discoveries of any importance in the offing, the restrictions were lifted (Smith 1986, p.288). A prospecting boom by private enterprise ensued. The first find in Ontario that brought attention was one made by Robert Campbell in 1948 at Theano Point along the east shore of Lake Superior, by following up on a report on the occurrence of coracite, a uranium mineral found by John LeConte (1847, p.173–175) in the 1800s. This discovery (the Camray) did not prove economic in itself. But it did arouse interest in the region, and indirectly led to the discovery of the Blind River–Elliot Lake uranium camp in 1953 (Hanula and Longo 1982, p.168–174). As reported by Smith (1986, p.288):

> The search [for uranium] now switched to the Bancroft area....Here in 1949, a local prospector Arthur H. Shore had found a uranium deposit in Faraday Township....In June, 1949, he organized Faraday Uranium Mines Limited, which, after an injection of American capital, began production in April 1957. Shore's discovery set off a staking rush that dwarfed the gold rush into Eldorado almost a century earlier. Before it was over, development work had been carried out on more than a hundred properties, and by the late nineteen-fifties there were four uranium mines operating at Bancroft.

In response to the uranium exploration in the province, E.W. Nuffield of the University of Toronto was commissioned in 1949–1951 to map the Montreal River area where the Camray discovery had been made. Also, E.M. Abraham, who had replaced Horwood on the Department's head office staff after the latter's untimely death, began a program of detailed township mapping of the Blind River–Elliot Lake area in 1954, a program which was carried on by J.A. Robertson and others into the 1960s and all but completely covered the area between Sault St. Marie and Sudbury (Robertson et al. 1969). In the period 1954–1956, J. Satterly also made a comprehensive study of the uraniferous pegmatite deposits of the Bancroft area, producing what some consider to be one of the Survey's classic geological reports (Satterly 1957).

In 1953, at the time of heated exploration in the Bancroft and Blind River–Elliot Lake areas, other prospectors focused their attention on base metals. A major discovery, the Geco, was made at Manitouwadge by 3 amateur prospectors from Geraldton, Roy Barker, William Davidowich and Jack Forster. The area was mapped by the author (Pye 1957) in the period 1954–1956.

In the 1920s, copper-lead-zinc deposits had been found within the Sudbury Basin. These deposits, were acquired and re-examined in the early 1950s by Ontario Pyrites Company Limited and found to be economic under the conditions of the time. Exploration by this company showed that the geological structure of the basin was more complicated than previously had been realized. Thomson (1957) began mapping of the basin in the period 1953–1954. After 2 seasons of field work, which was concentrated on the study of the Onaping Formation, Thomson (1957, p.3) "became convinced that a geologist with long experience in active and Tertiary volcanism could contribute much towards the solution of many perplexing fundamental problems of the Sudbury basin." He recommended that Howel Williams of the University of California be asked to act as a consultant on volcanology at Sudbury. Thomson stated:

> Dr. Williams was chosen because of his wide experience in volcanology in all parts of the world and his knowledge of depressed basins and pyroclastic rocks. In the summer of 1955 Dr. Williams came to Sudbury, and the author [Thomson] had an opportunity to show him the key localities of volcanic rocks in the area. Williams ensuing report (Williams 1956, p.57–89) is reproduced in full following this [Thomson's] report. The author considers it to be an important contribution to Sudbury geology and a landmark in Precambrian volcanology.

In 1954, the Department of Mines published its first Mineral Resources Circular on the copper, nickel, lead and zinc deposits of the province (Thomson et al. 1954). These reports or "catalogues" giving information on location, geological setting, dimensions and grades of deposits, and literature references have been largely supplanted by the files of the Geoscience Data Centre (*see* The Reorganization of 1972). They were, and still are, of particular use to prospectors who did (do) not have ready access to the files and the literature. One Mineral Resources Circular on gold deposits (Ferguson et al. 1971) is credited by Smith (1986, p.330–331) to have assisted in the recent discoveries of major gold deposits at Hemlo.

Industrial Minerals

In the 1950s, the attention of industry was not entirely directed toward metallic minerals. In 1951, at the eighth annual conference of Mines Ministers in Saskatoon, a Committee on Industrial Minerals reported: "It is believed a considerable fund of information pertinent to industrial minerals exists in the various provinces, but that such information is widely scattered, and in some cases is relatively inaccessible to the interested public." The Committee went on to recommend "that an effort be made to at least index, if not consolidate this information. To this end it is suggested that each Minister arrange to have his staff gather and compile as complete a listing as possible of published and unpublished technical reports concerning the industrial minerals of his province." The Committee's recommendation was accepted by the ministers of the day.

In Ontario, the task was assigned to the late D.F. Hewitt, who enlarged the terms of reference to include field examinations, investigation of methods of treatment and recovery, and compilation of data on uses, specifications and markets (Figure 2.14). The first report, entitled "Asbestos in Ontario" was published as Industrial Minerals Report No. 1 (Hewitt 1953), starting a long series of highly popular reports, totalling 42 in number by 1975. In 1952–1953, Hewitt examined and reported on the kyanite deposits of the Sudbury and Nipissing districts, graphite in Frontenac County, and muscovite in the Parry Sound District. In 1954–1956, Hewitt reported on the nepheline, fluorspar and corundum deposits of the province. In 1958–1959, he turned his attention to the stratigraphy, distribution and economic utilization of limestones, and in 1960, to the rock salt deposits of southwestern Ontario.

In the mid-fifties, the space-age metal lithium came into prominence as a potential contributor to rocket fuel and as a means of widening the range of fluidity of lubricating oils. At the 1955 annual convention of the Prospectors and Developers Association, the lithium deposits of the Preissac–La Corne area in Quebec were described by Latulippe and Ingham (1955). Samples of spodumene were on display. Years earlier, E.W. Hadley of Auden, Ontario, had discovered a pegmatite deposit at Georgia Lake south of Beardmore. The pegmatite contained an unusual prismatic mineral which he could not identify. At the convention he noted that the spodumene on display was similar, whereupon he contacted Gorden Miller of Conwest Exploration Company Limited. Impressed by the occurrence, Miller submitted samples to the author for identification. The author, in turn, presented Miller's samples to H. Quackenbush,

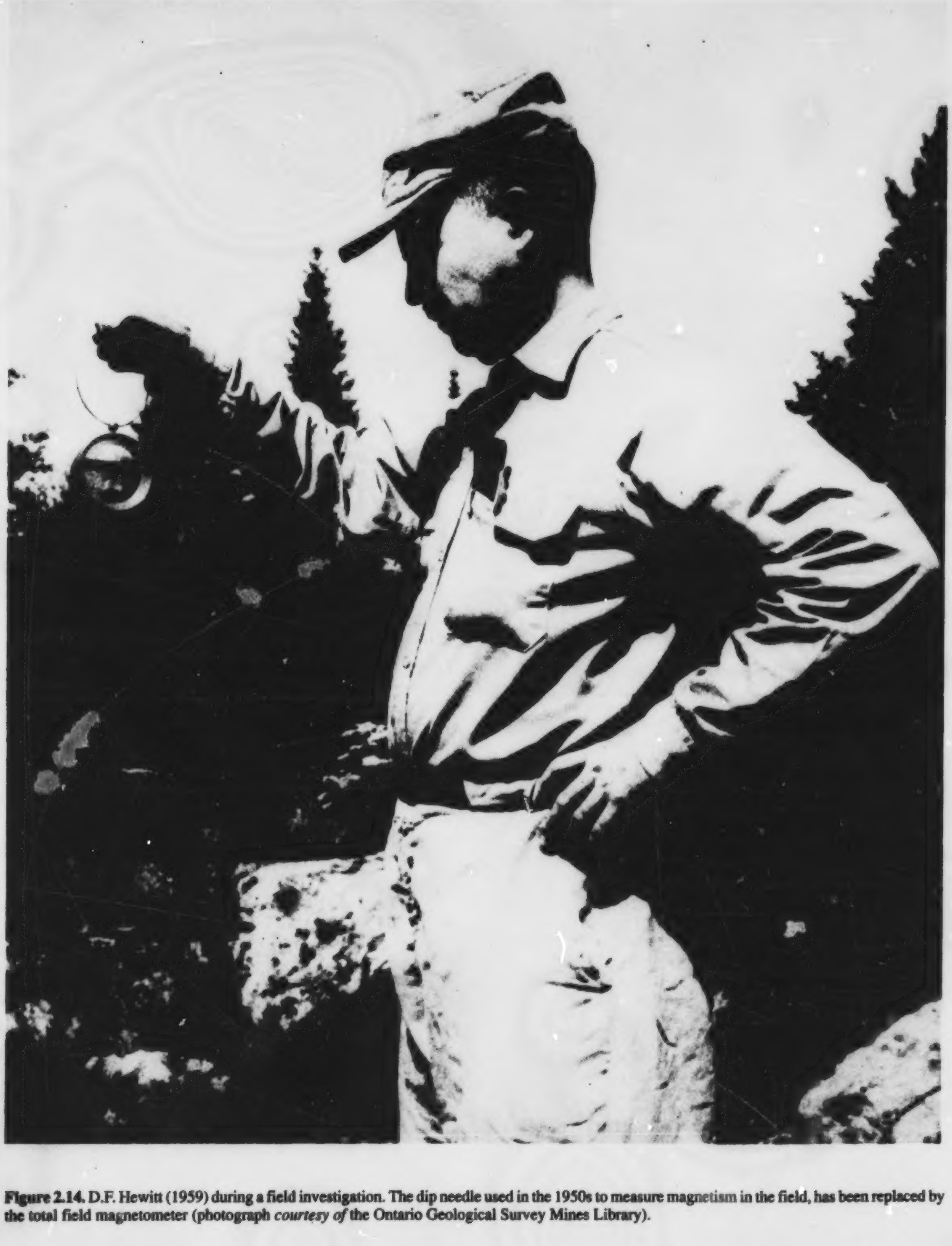

Figure 2.14. D.F. Hewitt (1959) during a field investigation. The dip needle used in the 1950s to measure magnetism in the field, has been replaced by the total field magnetometer (photograph *courtesy of* the Ontario Geological Survey Mines Library).

a Fort William dentist who, using a spectroscope he had made himself, confirmed that the mineral was indeed spodumene. Miller proceeded to stake the deposit for his company. As news of the discovery was publicized, a staking rush ensued, and within a short time numerous additional lithium deposits were located. Two deposits were found on diamond drilling by Nama Creek Mines Limited to contain over 4 million tons of material considered by the company to be of commercial grade, and underground development was initiated. The author mapped the area in the period 1956–1959 (Pye 1965). It is unfortunate that, because of a poor market demand, none of the deposits in the area have been brought into production.

Groundwater Resources

Groundwater resources normally are not considered to be the responsibility of the Ontario Geological Survey. However, in 1945, an organizational unit was set up under A.K. Watt to collect basic hydrologic data and to conduct geological and groundwater surveys in areas overlain by glacial and other surficial deposits. The staff and all the records of the unit, with the exception of those relating to the systematic mapping of Pleistocene deposits, were transferred to the Ontario Water Resources Commission on April 1, 1957.

Airborne Magnetometer Surveys

Following the success of the aeromagnetic survey of the Marmora area, additional airborne surveys were made in a number of localities. In 1954, a combined magnetometer and scintillometer survey was made by Aeromagnetic Surveys Limited of 12 townships in the vicinity of Blind River; a similar survey of 36 townships in the same area was made in 1955; and a survey of 21 townships in the vicinity of Bruce Mines in 1956. In 1958, 2 airborne magnetometer surveys were made for the Department of Mines. One, by Aeromagnetic Surveys Limited, covered 18 townships (1070 square miles) south of Kapuskasing; the other, by Spartan Air Services, covered an area of approximately 310 square miles south of Sudbury, including part of the Whitefish Lake Native Community.

Finally, In the last year of the decade, in connection with the federal government's "Roads to Resources" program, the Ontario Department of Mines entered into an agreement with the Department of Mines and Technical Surveys at Ottawa for a contact with Spartan Air Services to conduct an airborne magnetometer survey of an area of 60 000 square miles in the northwest part of the province. This survey involved 120 000 line miles of flying at 1/2 mile intervals (Figure 2.15a).

Ontario's Mineral Map

A publication highlight of the 1950s was Ontario's first Mineral Resources Map produced jointly by the geological and cartographic staff (Ontario Department of Mines 1953). This map shows the locations and types of Ontario's mineral deposits. Its particular value lay in its simplified portrayal of the variety and wealth of the province's total mineral heritage in relation to geology for the investor and developer in the mineral sector. J.E. Thomson, the originator of the map was habitually kidded by the cartographers for devising the shortest of all the Department's geological map legends. The map was produced in 3 colours: green for volcanic rocks; grey for sedimentary rocks; and pink for intrusive rocks. Later editions of the map showed a greater complexity.

THE 1960s: A TIME OF EXPANSION AND REORGANIZATION

In the 1930s, much of the mapping done by the Ontario Department of Mines was of a reconnaissance nature for publication at scales of 1 inch to 1 or 2 miles (1:63 360 or 1:126 720)). But in the late 1940s and early 1950s, much of the mapping became very detailed, for publication at a scale of 1 inch to 1000 feet (1:12 000), to serve the anticipated needs of the gold prospector. Then, with the increasing availability of good quality airphotos, and the realization that township mapping at this scale would make adequate coverage of the known greenstone belts an interminable task, mapping at 1 inch to 1/4 mile (1:15 840), for final publication at 1 inch to 1/2 mile (1:31 680), came into vogue. At this time, heavy dependence was placed on the seasonal employment of academic personnel to round out the mapping program; and by the late fifties, about 15 parties were placed in the field annually. But even at this rate, adequate coverage of the province still seemed a long way in the future (Figure 2.15b).

In 1960, the late J.E. Thomson pointed out that at the current rate of mapping it would take another 25 years to complete the mapping required, at the standard of the time. Further, experience in the late fifties had demonstrated that professors, with the increasing availability of research funds, were becoming less interested in field mapping as a research function, were becoming increasingly difficult to recruit, and in the interests of academic freedom, no longer could be relied upon to adhere to standards, specifications and production schedules imposed by a government agency. The Department successfully petitioned Treasury Board for permission to recruit 25 additional staff geologists over a period of 5 years. As part of this expansion, Paul Karrow was added to the staff in 1960, to be followed in 1961 by S.A. Ferguson, A.M. Goodwin, P.E. Giblin and J.A. Robertson; and in 1963, by the present Director of the Ontario Geological Survey, V.G. Milne, and others, bringing the Toronto staff to 12.

The Reorganization of 1961

After D.P. Douglass succeeded Harold Rickaby as Deputy Minister in 1961, the Department of Mines was reorganized in the light of the anticipated expansion of the Geological Branch. It was at this time that the title of the Provincial Geologist, then M.E. Hurst, was changed to "Director", and J.E. Thomson became "Assistant Director", with the less formal title of "Chief Geologist" more in common with the practice of industry, rather than that of a government bureaucracy. The Chief Geologist was made responsible for all geological mapping, while the Director, the former Provincial Geologist, retained responsibility for the Resident Geologists in the field offices. The position of "Senior Geologist" also was created. The publication of all maps and reports became the responsibility of J. Satterly; and cartography, editorial services, statistics, the Mines Library, and the Publications Distribution Office were placed under his control as head of a "Division of Maps and Reports". At the same time, D.F. Hewitt was appointed "Chief" of a new "Industrial Minerals Section". To complete the reorganization, the Provincial Assay Office in Toronto and the Timiskaming Testing Laboratory in Cobalt were joined in a new "Laboratories Branch" under the former "Provincial Assayer", D.A. Moddle, as Director (Figure 2.16a and 2.16b).

Throughout the early sixties, the Survey continued to expand. By 1966 it boasted a permanent staff of 28 geologists. In 1962, P.E. Giblin became the first Resident Geologist at Sault Ste. Marie, and the following year saw another resident geologist's office opened in Sudbury under the direction of K.D. Card, bringing to 7 the number of such offices located strategically across the province.

The Reorganization of 1966

Following the retirement of M.E. Hurst in 1966, and the appointment of J.E. Thomson as Director, the Branch experienced a second reorganization.

A "Geological Surveys Section", with responsibility for all field mapping was placed under the author as "Chief Geologist"; a "Resident Geologist's Section", initially under the Director, and later in 1968, under G.R. Guillet, was set up; J. Satterly became "Senior Precambrian Geologist" with responsibility for the critical review of all maps and reports before these were passed to a "Publications Production Section" under J.A. Haddon, Chief Cartographer. This section replaced the former "Division of Maps and Reports", but now included a "Preliminary Map Unit" under

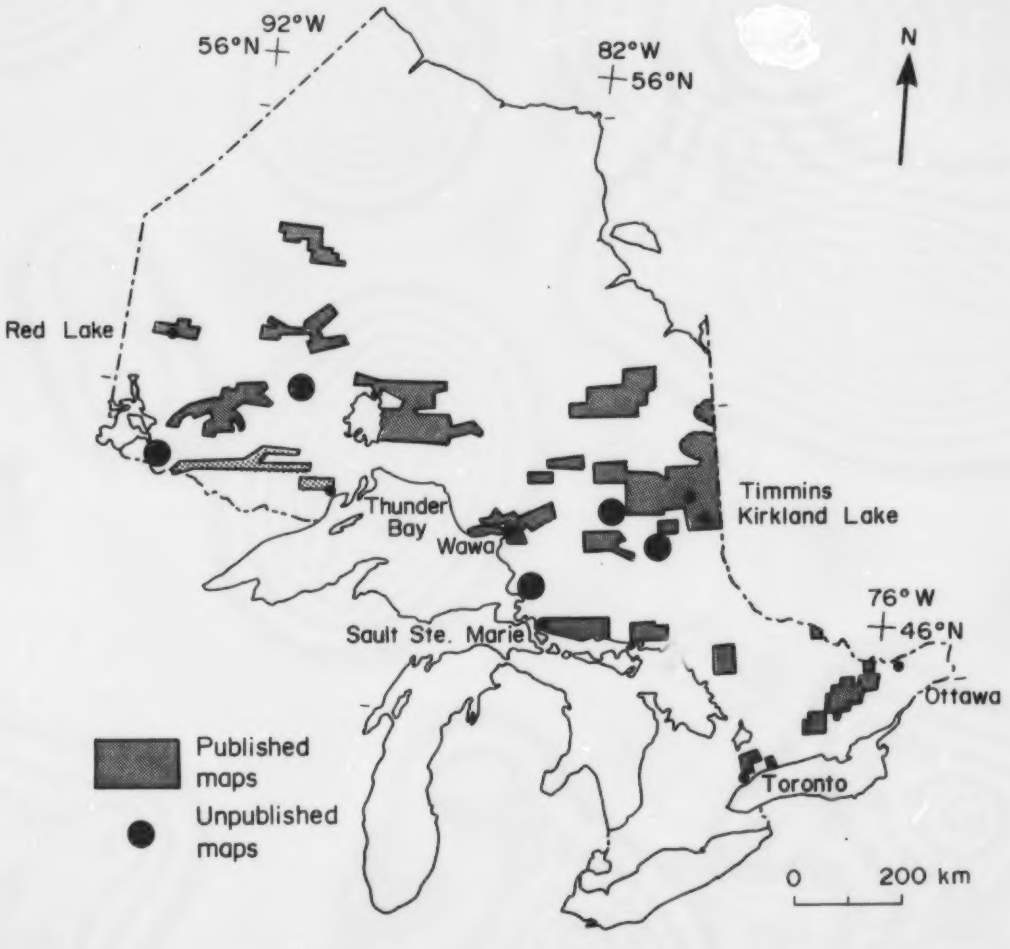

Figure 2.15a. A sketch map showing the areas of Ontario covered by airborne magnetic and electromagnetic surveys funded partly or completed by the Ontario government to 1989 (*compiled by* the Geophysics/Geochemistry Section).

R. Balgalvis, to prepare the geologists' rough field maps for public release in uncoloured format, as soon as possible after each field season. This initiative was soon to be copied in other Canadian jurisdictions.

SELECT COMMITTEE ON MINING

The expansion and reorganization of the Ontario Department of Mines was given another boost with the publication of the report by the Select Committee of the provincial legislature on Mining. On May 8, 1964, the Legislature had appointed a committee of 13 of its members:

> ...to inquire into and report upon methods of stimulating prospecting and mining exploration and development in Ontario and, without limiting the generality of the foregoing, such matters as the services available to the mining industry, regulations governing the financing of mining prospecting and development, the effects of mining taxation on the growth of the industry, health measures in the mining industry, and any other related factors the study of which would bring about renewed interest and activity in that industry and thus strengthen the economic position of the northern communities and Ontario as a whole.

This Committee, under the chairmanship of Rene Brunelle, invited submissions from all interested parties, and from October 14, 1964, to January 6, 1966, held 22 public hearings throughout the province.

In regard to the Geological Branch, the Committee recommended: the establishment of a program for the classification, storage and rapid retrieval of geological and exploration data for the province's mineral deposits; expansion of its program of prospectors' classes, which had been on the descendancy since the retirement of W.D. Harding in 1959 (Pye 1989); and inclusion of geophysicists and geochemists on the staff. In keeping with these recommendations, an "Information and Education Section" made up of the Mines Library (Figure 2.17a and 2.17b), a data centre, and an "Education" unit was set up under the direction of Stewart A. Ferguson. Under Ferguson's initiative, the prospectors' classes were increased to an average of

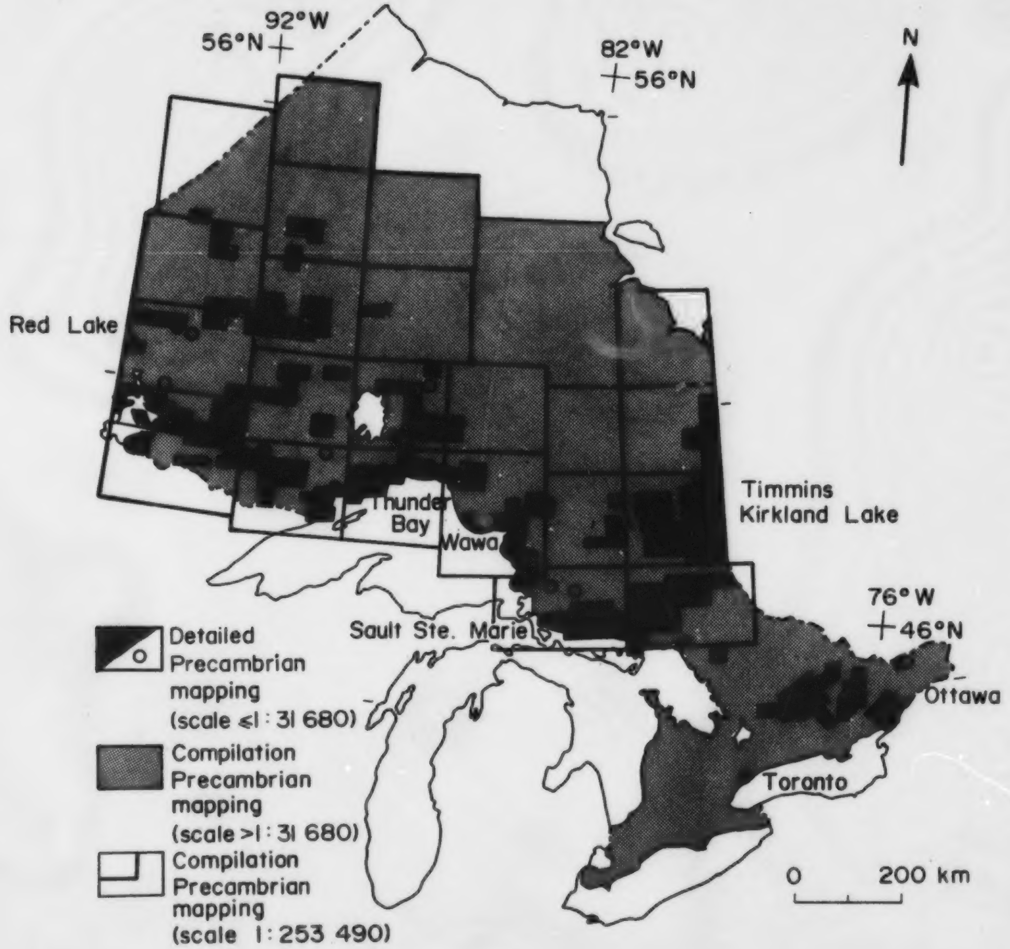
92°W
56°N
82°W
56°N
N
Red Lake
Thunder Bay
Wawa
Timmins
Kirkland Lake
76° W
46°N
Sault Ste. Marie
Ottawa
Toronto
0
200 km
Detailed Precambrian mapping (scale ≤1 : 31 680)
Compilation Precambrian mapping (scale >1 : 31 680)
Compilation Precambrian mapping (scale 1 : 253 490)

Figure 2.15b. A sketch map showing the areas of Ontario covered by detailed mapping (scale 1 inch to 1/2 mile [1:31 680] and less), and compilation mapping (scale greater than 1 inch to 1/2 mile [1:31 680] and including 1 inch to 4 miles [1:253 440]), completed by the Ontario Geological Survey up to 1989 (*compiled by* G.W. Johns, Precambrian Geology Section).

11 classes per year during the period 1968–1975 (*see* Figure 2.4).

At first, it proved difficult to recruit a suitable geophysicist or geochemist. Not until 1969 did R.S. Middleton join the Geological Surveys Section as the Survey's first full-time geophysicist, and W.J. Wolfe as the first full-time geochemist.

REVIEW AND RESOURCES SECTION

Until the mid 1960s, the Survey relied on a single employee for its editorial work. Then, in 1965, 3 additional editors were hired in expectation of a considerably increased production from the expanded geological staff. None of the editors, however, had any geological training or experience. As a result, it was common to find the geologist's meaning distorted after a sentence or paragraph had been rephrased in the interests of good grammar. This situation failed to improve with time. Eventually, to remedy the problem, the unit was disbanded and the discharged editors replaced by professional geologists. Also at the time, the cartographers of the Division of Maps and Reports began to regard with disdain the Branch's uncoloured preliminary maps, and conflict developed with the Preliminary Map unit.

In response to the many problems, the Cartography unit, under J.A. Haddon, was given Section status; and the editorial services function, Preliminary Map unit, and mineral deposit inventories were added to J. Satterly's other responsibilities in a new "Review and Resources" Section. This Section came into being in 1967, when G.M. Scott was appointed as the first "Review Geologist" and A.G. Sergiades became the first "Resources Geologist".

A large backlog of unedited reports had built up because of the expansion which began in 1961. To obviate serious delays in publication, it was at this time that the present practice of Open File releases was initiated, so that the new information in the geologists' unedited manuscripts could be made available to the interested public quickly after final reports were submitted.

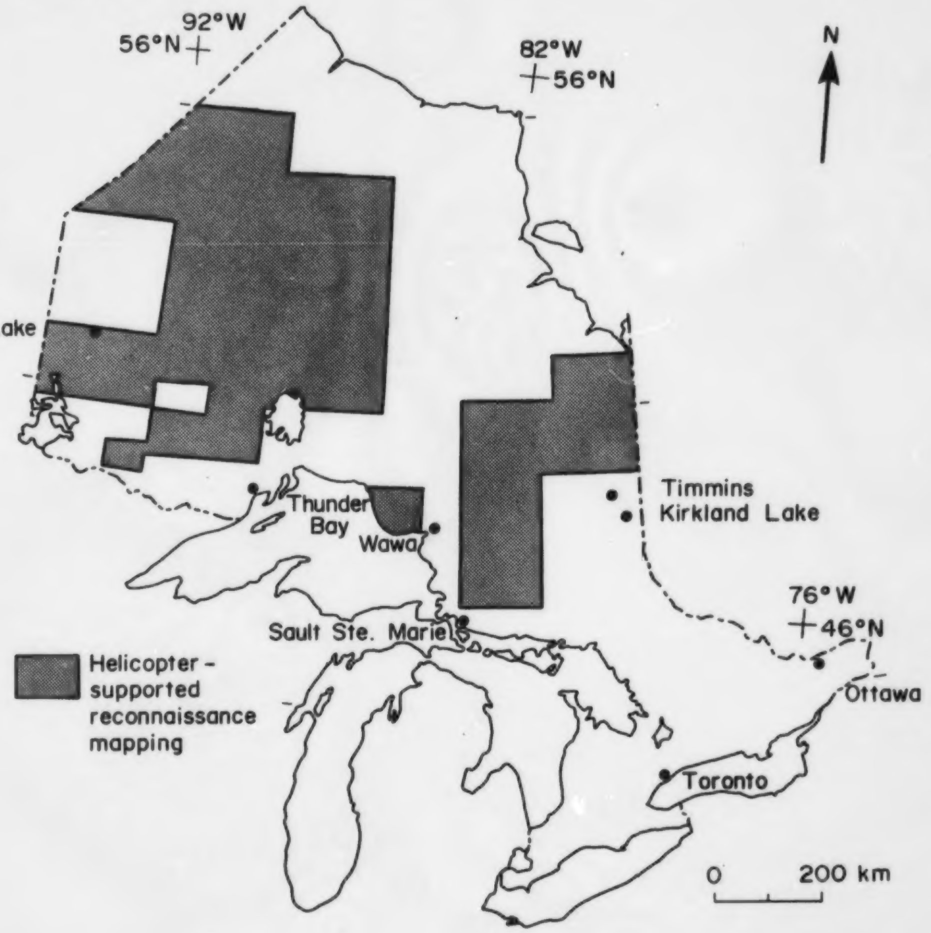

Figure 2.15c. A sketch map showing the areas covered by helicopter-supported regional reconnaissance surveys between 1966 and 1976 (*compiled from* Ontario Geological Survey Summaries of Field Work).

Geological Compilation Maps

Early in the 1960s, Hunting Air Services Corporation commenced preparation of geological compilation maps, using the results of airborne magnetometer surveys flown for various companies and data made available by the Ontario Geological Survey. These maps proved very useful as a planning aid for the explorationist. However, their market price was high, and beyond the reach of the ordinary prospector. The Department itself felt that its budget did not permit the purchase of these maps. Perhaps with this in mind, J.E. Thomson divided the map of the province into 22 areas of roughly equal size, with boundaries determined in regard to the known geology rather than any particular system of geographic reference lines, and initiated a program of geological compilation, for final publication at a scale of 1 inch to 4 miles (1:253 440), a program which fell mainly on the shoulders of the Resident Geologists. The first compilation map, the Timmins–Kirkland Lake sheet, was published in 1963. Other sheets followed, at a rate of one or two per year until the program, except for revisions as new information has become available, was completed over the following decade (*see* Figure 2.15b).

Helicopter-Supported Regional Reconnaissance Surveys

In the preparation of the Compilation Series maps, it soon became apparent that there was a dearth of information in the extreme northern parts of the province. Going back in time, the Geological Survey of Canada had initiated helicopter-supported reconnaissance surveys in the Northwest Territories with "Operation Keewatin" in 1952, and was thus able to map an area of 57 000 square miles in a

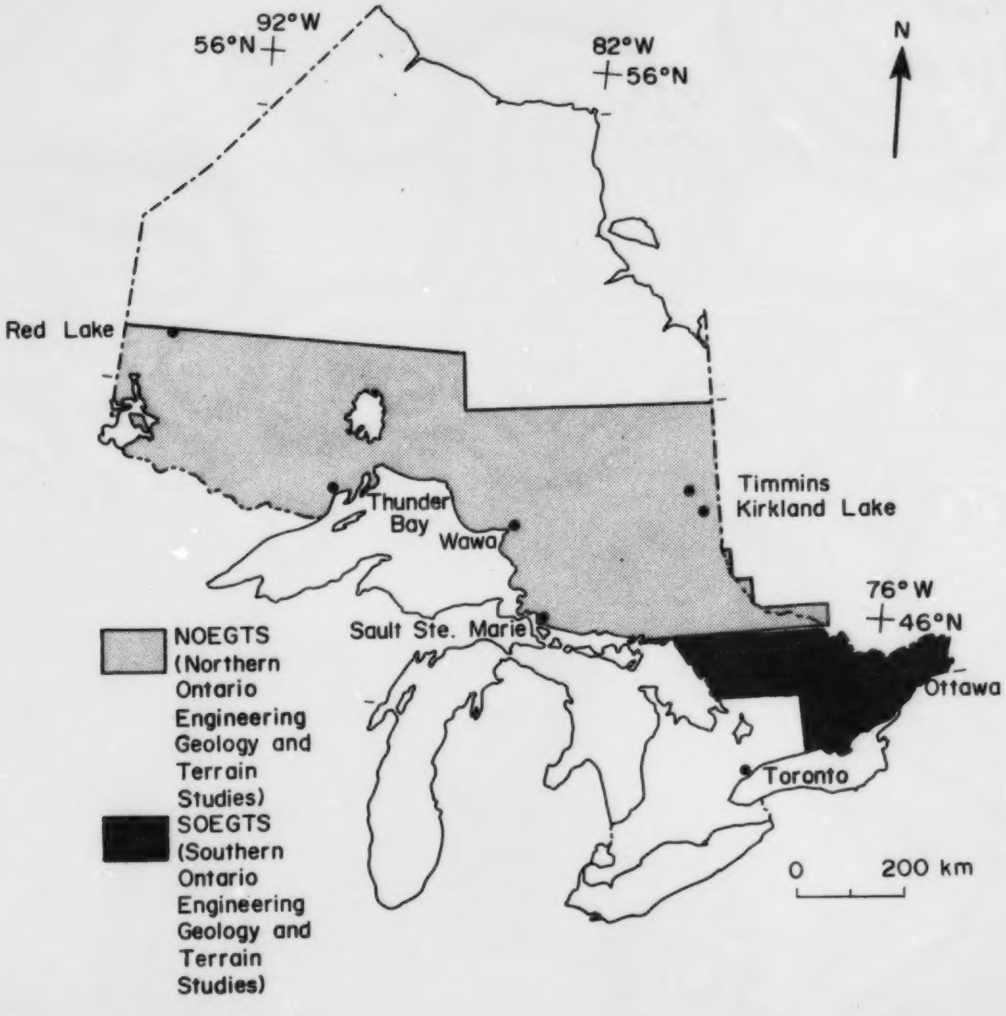
92°W
56°N
82°W
56°N
N
Red Lake
Timmins
Kirkland Lake
Thunder Bay
Wawa
76° W
46°N
Sault Ste. Marie
Ottawa
Toronto
0
200 km
▨ NOEGTS (Northern Ontario Engineering Geology and Terrain Studies)
■ SOEGTS (Southern Ontario Engineering Geology and Terrain Studies)

Figure 2.15d. A sketch map showing the areas of Ontario covered by NOEGTS/SOEGTS (Northern/Southern Ontario Engineering Geology Terrain Studies), from the beginning of the program in 1978 to 1981. New SOEGTS have since been completed for southern Ontario but have yet to be released (*compiled by* the Engineering and Terrain Geology Section).

Figure 2.16a. Ontario Geological Survey Laboratories, located on Grenville St., Toronto, 1989. J. Tsigaris analyzing samples in the X-ray Laboratory (photograph *courtesy of* E.B. Freeman, Geoservices Section).

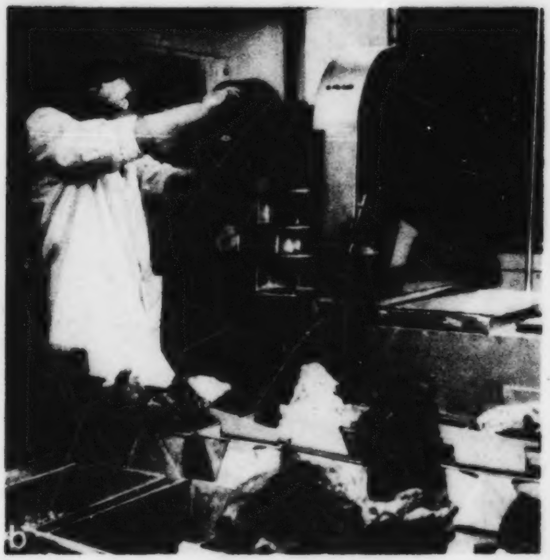

Figure 2.16b. Ontario Geological Survey Laboratories, located on Grenville St., Toronto, 1989. D. Ross crushing rock samples in the Assay Laboratory (photograph *courtesy of* E.B. Freeman, Geoservices Section).

single field season (Hanula and Longo 1982, p.71). The Ontario Department of Mines started a similar program in 1966 with "Operation Kapuskasing" (Bennett et al. 1967) (Figure 2.18). This was followed by other helicopter-supported surveys until the reconnaissance mapping of the province was completed in 1976 (*see* Figure 2.15c).

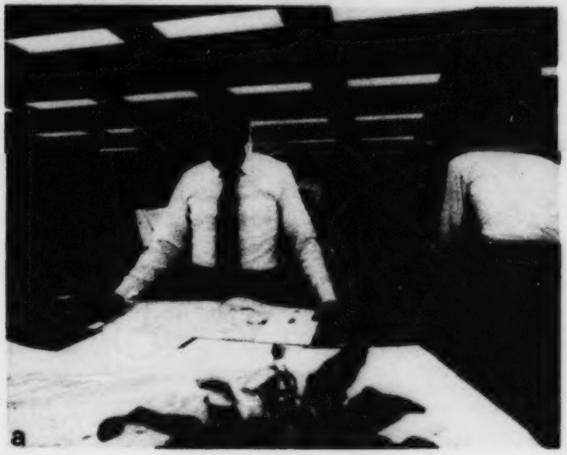

Figure 2.17a. Ontario Geological Survey Mines Library (1989). The Mines Library provides the Ontario Geological Survey staff and the public with access to published geological documents from around the world.

Figure 2.17b. Ontario Geological Survey Mines Library (1989). L. Vanasse–Murphy stands at plan files of current date claim maps. Shelves store assessment file records of work performed by various mining companies in Ontario (photographs *courtesy of* E.B. Freeman, Geoservices Section).

In 1946, J. Satterly prepared the first complete geological map of the province. This was then published at a scale of 1 inch to 20 miles (1:1 267 200) and was revised by J. Satterly in 1958. However, the compilation maps, and the helicopter-supported reconnaissance surveys, resulted in Satterly's map becoming outdated. At this time, V.G. Milne, L.D. Ayres, S.B. Lumbers and D.W. Robeson (cartographer) were given the task of preparing a new map. The updated map was published in 1972 as 5 sheets, at a scale of 1 inch to 16 miles (1:1 013 760).

Airborne Magnetometer Surveys

Publication of the results of the "Roads to Resources" airborne magnetometer survey (*see* The Exciting 1950s

Figure 2.18. Department of Mines geologist P.C. Thurston studies an outcrop during the helicopter-supported regional reconnaissance survey of the Fort Hope area, 1969 (photograph *courtesy of* the Ontario Geological Survey Mines Library).

Airborne Magnetometer Surveys) commenced on March 30, 1960, when 10 maps, at a scale of 1 inch to 1 mile (1:63 360) were issued. By year's end, 120 maps, out of an anticipated total of 160, had been released. In May, 1961, an agreement was made with the Federal Department of Mines and Technical Surveys to share the cost of an additional survey of an area of 35 000 square miles, south of the area flown in 1959–1960. This survey required 80 950 line miles of flying at intervals of approximately 2250 feet with a terrain clearance of 1000 feet.

The maps yielded by the 2 surveys were found to be quite useful by industry and were well received and acclaimed to be a positive aid to mineral exploration. Early in 1962, discussions were held with the Department of Mines and Technical Surveys in regard to a joint program toward completing aeromagnetic coverage of northern Ontario, at a rate of 28 000 to 35 000 square miles per annum, within 7 years. The work was actually completed by 2 contracts, 1 awarded to Spartan Air Services, the other to Hunting Air Services, by 1967, within 5 years.

THE 1970s AND THE MINISTRY OF NATURAL RESOURCES

At the close of the 1960s, following a decade of expansion and the initiation of several well-received new programs supplementing regular geological mapping, the Branch was left with a feeling of confidence in its capabilities and an augury of well-being for the seventies. Political events, however, were to change things to a completely unexpected degree. Following the recommendations of a Committee on Government Productivity, which had been deliberating on government performance for a year or two, the Premier, William Davis, in late 1971, announced plans to reorganize the provincial government to reduce the number of Cabinet Ministers and streamline the administration in the interests of increased efficiency. Included in these plans, to become effective April 1, 1972, the former Department of Mines and Northern Affairs and the Department of Lands and Forests were to be amalgamated to form a new Ministry of Natural Resources. At this time, the Geological Branch, the Mines Inspection Branch, the Laboratories Branch (recently renamed, the Mineral Research Branch), and a new Mineral Resources (policy) Branch were united in an Ontario Division of Mines (Figure 2.19), while the former Mining Lands Branch became part of a Division of Lands.

The Reorganization of 1972

Within the major restructuring, the Geological Branch itself underwent several changes. In the months preceding the 1972 reorganization of government, the Geological Branch was approached by the former Department of Highways for mineral potential or capability analyses in the form of maps of a large part of northwestern Ontario to aid in highway location to provide optimum access to the province's natural resource endowment. At the same time, there was recognition of the need for metallogenic studies by staff geologists beyond the confines of specific map areas. Also, because of staff turnover, the Branch continued to depend on the Geological Surveys Section for mineral deposit inventories as a part-time activity. With the help of the Department of Highways, the Treasury Board of Cabinet was approached, and approval was obtained to create a Mineral Deposits Section (Robertson 1975) with a staff of 9 geologists only 5 were provided by the new Ministry.

On April 1, 1972, the Geological Branch emerged from the reorganization of government with a structure quite different than was created in 1966–1968:

1. The new Mineral Deposits Section came into being under the direction of J.A. Robertson.
2. The Cartography Section was removed from the Geological Branch and merged with a similar unit of the Surveys and Mapping Branch of the Division of Lands.
3. The Mines' Library fell under the control of a new Policy Research Branch, the objective being to merge it with the pre-existing Lands and Forests' Natural Resources Library.
4. The Resident Geologists Section was disbanded and the Resident Geologists came under the control of the Ministry's regional offices in which they were located.
5. The Review and Resources Section and the Data Retrieval and Education Section were wound up, and a Scientific Review (editorial) Office, a Geoscience Data Centre and a Geoscience Information Office

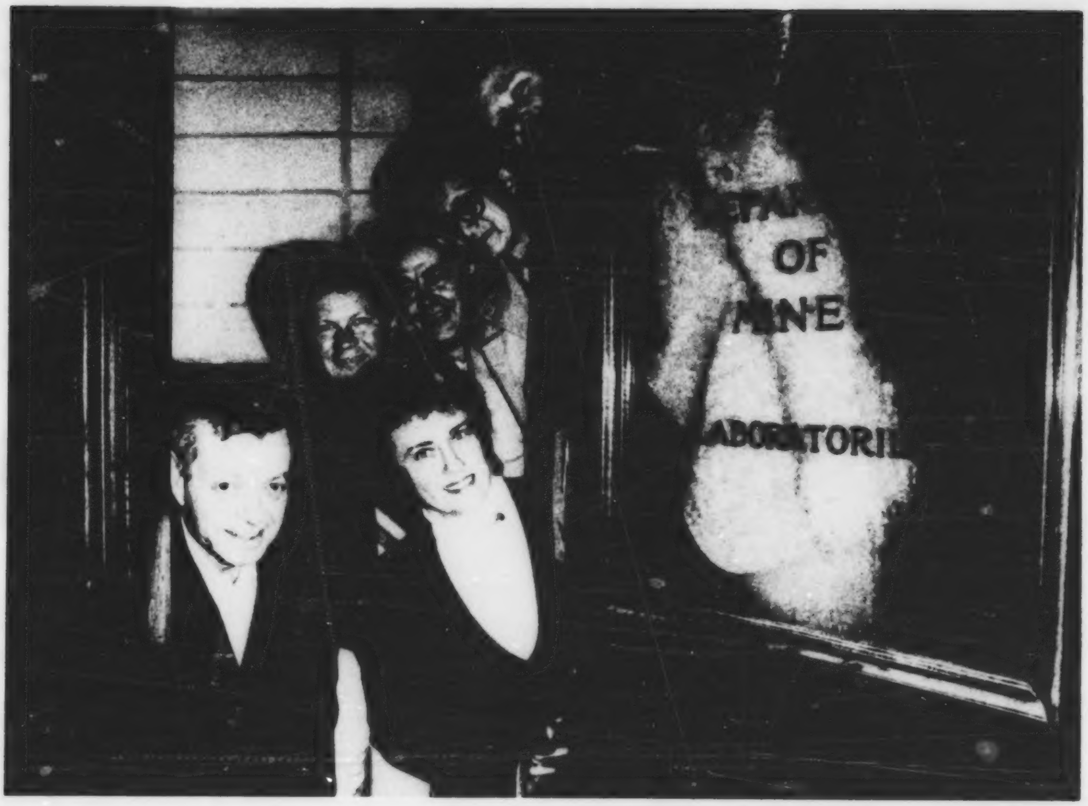

Figure 2.19. Department of Mines Laboratories situated in the Whitney Block, Queen's Park, Toronto (bottom to top: A. Lane, P. Fyfe (née Barchuck), R. Smith, R. Laakso, J. Hwang and E.B. Freeman). In 1972, the Department of Mines became the Ontario Division of Mines and moved their location to Grenville St., Toronto, with the OGS following suit in the mid-1970s (photograph *courtesy of* the Ontario Geological Survey Mines Library).

responsible for mineral exploration (Prospectors) classes and for disseminating information on the geology and mineral deposits of the province to the lay public, were transferred to a new Geoservices Section under C.R. Kustra.

6. The administration of the Mineral Exploration Assistance Program initiated by Leo Bernier, Minister of Natural Resources in 1971, became the responsibility of S.A. Ferguson as head of a Special Projects Section.
7. A second Industrial Minerals Section was created in the Mineral Resources (Policy) Branch. To avoid confusion, the name of the Industrial Minerals Section in the Geological Branch was changed to Phanerozoic Geology Section, a name which proved unsatisfactory to senior management and was later changed to Engineering and Terrain Geology Section. Later in the decade, as responsibilities widened, the Engineering and Terrain Geology Section was subdivided into Paleozoic and Quaternary subsections and an Aggregate Assessment Office.
8. The Geological Surveys Section, now under the supervision of V.G. Milne, was renamed the Precambrian Geology Section, recognizing that surveys of Cambrian and younger terrain, as they had for several years, would continue to be the responsibility of the Engineering and Terrain Geology Section.
9. The Precambrian Geology Section itself was subdivided into subsections, each with its own supervisor: North Archean, Central Archean, South Archean and Proterozoic, recognizing the major tectonic units of the Shield in Ontario.
10. Finally, to recognize the growing importance of geophysics and geochemistry in the interpretation of Ontario's geology and mineral exploration, a Geophysics/Geochemistry Section (Figures 2.20a and 2.20b), was established under the supervision of W.J. Wolfe.

In the mid 1970s, declining tax revenues put ever-increasing pressures on the Ministry to, in the common expression of the time "do more with less". The new Ministry was forced to undergo a second reorganization in 1977 because of decreased budget allocations. Following this in 1978, the Mineral Research (Laboratories) Branch and the Geological Branch were joined under the directorship of the author as the Ontario Geological Survey (Pye 1985, p.39–40).

Figure 2.20a. K. Steele sorting and bagging overburden samples during a 1985 field investigation for analysis by the Geochemistry Section (photograph *courtesy of* D. Wadge, Geophysics/Geochemistry Section).

Figure 2.20b. Taras Dusanowskyj of the Geophysics Section uses a portable graviometer at a local base station in northwestern Ontario (photograph *courtesy of* D. Wadge, Geophysics/Geochemistry Section).

The 1970s were a difficult time for the Geological Branch or Survey, as at first, pressures were exerted by the Regional Directors and District Managers to effect a decentralization of the professional staff (Pye 1986a and 1986b), as had happened to the Resident Geologists in 1972; then, in the mid-seventies, serious discussion centred around transferring the Mineral Deposits Section to the Mineral Resources Branch and combining the Geoservices Section, the Mineral Research Branch, and the Statistics Section of the Mineral Resources Branch to form a Mineral Services Branch (Pye 1987a). In the late seventies, consideration was also given to "privatizing the Survey" as a selfsupporting service (Pye 1987b). Despite these problems, exacerbated as they were by serious budget and staffing constraints, the 1970s were not without their accomplishments.

Stratigraphic Mapping of the Archean

During the 1960s, the Survey's attention had been diverted away from the Timmins–Kirkland Lake area because of developments elsewhere. However, the discovery of the Kidd Creek orebody and other base metal deposits was soon to refocus the Survey's attention on this important part of the province. In the 1970s, a renewal of township mapping in the area (for the now-customary 1/2 mile, (1:31 680) publication scale), was early seen as leaving something to be desired in regard to the recognition, tracing and correlation of geological units. In 1975, a program of compilation of all available information supplemented by field mapping of designated NTS quadrangles for publication at 1 inch to 1 mile (1:63 360), was initiated. This program, termed "synoptic mapping" was started by D.R. Pyke (1975, p.2) in the Timmins–Matachewan part of the area (Figure 2.21), and by L.S. Jensen in the Kirkland Lake part. This work, following recognition of the komatiite suite of volcanic rocks by Pyke (Pyke et al. 1973), and by the use of geochemical methods to classify volcanic rocks of the komatiite, tholeiite, calc-alkalic and alkalic suites by Jensen (1976), has done much toward working out the stratigraphy and structure of this part of the Abitibi Subprovince, presenting a new and extremely valuable synthesis and framework for future mineral exploration, particularly for both gold and massive sulphide deposits. The successful development of stratigraphic mapping of "greenstones", using geochemical techniques, is now widely recognized as having been a major contribution to the current understanding of the Precambrian geology of the province. While in the 1960s and earlier, little attempt was made to subdivide Archean "greenstones", the practice of stratigraphic mapping introduced by Pyke and Jensen became common in the Survey's mapping projects.

Geochronology

In the early 1970s, there was recognition that a sound program of geochronology could be a valuable tool in unravelling the Precambrian history of the province, and this in turn could lead to a better understanding of the origin and localization of mineral deposits. John Wood of the Precambrian Geology Section was given the task in 1973 to study the "state of the art" and make recommendations in regard to establishing a geochronology laboratory in Ontario to supplement and assist the regular mapping program. John Wood presented an excellent report outlining the advantages of such a program, and pointed out strongly that such a laboratory should be fully equipped for both U-Pb and Rb-Sr dating methods. He recommended that any new laboratory

Figure 2.21. Field assistants to D.R. Pyke involved in the Precambrian mapping investigation of the Timmins area of the Abitibi Subprovince, 1973 (left to right: J. Lorsong, T.L. Muir, B. Seabrook, R. Hillier and Bolero). This study was one of several of the early synoptic mapping programs (Pyke 1982).

in Ontario should be research oriented rather than production oriented. Staffing constraints at the time made it very unlikely a geochronology laboratory could be established within the Ontario Geological Survey. In the meantime, S.B. Lumbers of the Department of Geology and Mineralogy of the Royal Ontario Museum had also expressed a need for a new geochronology laboratory using Zr-dating methodology developed by T.E. Krogh at the Geophysical Laboratory in Washington D.C. At the author's request, K.D. Card, Chief, Geophysics/Geochemistry Section and John Wood, Precambrian Geology Section, together with S.B. Lumbers of the Museum, drafted a proposal for Branch/ Ministry submission to the Cabinet Committee on Resources Development recommending the establishment at the Museum of a geochronology laboratory similar to that in Washington. On June 9, 1975, Leo Bernier, Minister of Natural Resources, announced that a $300 000 grant was being made to the Museum and that construction of the facilities would start immediately. The Jack Satterly Geochronology Laboratory, under the leadership of T.E. Krogh, was officially opened by Frank Miller, who had succeeded Leo Bernier as Minister of Natural Resources on March 9, 1977 (Pye 1980, p.2). The laboratory is named after Jack Satterly (former head of the Review and Resources Section), a geologist noted for his attention to detail, accuracy and high standards of quality (*see* Figure 2.13).

The Survey shares in the operating expenses and supports a resident geochronologist at the Laboratory, guaranteeing priority access. Any reader of the geological literature on the Precambrian geology of the province will know that the Laboratory has been extensively used to great advantage by the Survey. The precise dating of rocks has been the key to many of the advances that have been made in the understanding of rock forming processes and mineralization in the Canadian Shield. Results of this work are reviewed by Corfu and Davis (this volume).

Federal–Provincial Uranium-Reconnaissance Program

In 1975, the Ontario Ministry of Natural Resources entered into an agreement with the Geological Survey of Canada to jointly fund a systematic uranium-reconnaissance survey of the province. This program consisted of complementary airborne radiometric (Figure 2.22), and lake-sediment geochemical surveys. The purposes of the surveys were primarily to supply the mining industry with exploration data that might lead to the identification of areas where there could be a high probability of finding mineable uranium deposits, but which could also prove useful for sound mineral management practices by government. The program started with an airborne gamma-ray spectrometer survey in the extreme northwest corner of the province, and then continued back and forth across the province, in the form of either airborne spectrometer surveys or geochemical surveys, depending on the geology and the nature of the terrain until, in 1979, all areas of interest had been covered. There is little evidence that the program was successful in the location of previously unknown economic uranium deposits. Perhaps, this justifies the federal government's decision to terminate its participation in the program in September, 1978. The geochemical data, however, have been valuable in private sector exploration for base metal deposits, and were useful for the provincial government's landuse planning exercises of the 1970s and early 1980s.

Aggregate Resources Inventory

Very early in the 1960s, D.F. Hewitt (1962, p.2–3) warned of the danger of restrictive ordinances imposed by the various municipalities in southern Ontario on the mining of construction aggregates because of environmental concerns, and the effects of these ordinances on supply and cost. The concerns of the municipalities were exacerbated by the burgeoning construction activity of the 1960s in the Toronto area. In 1975, the Ontario government appointed an "Aggregate Working Party" under the chairmanship of George A. Jewett, Executive Director, Division of Mines, Ministry of Natural Resources, to investigate aggregate mining in the province and to recommend ways in which the provincial objective of ensuring the ready availability of low-cost aggregates could be reconciled with the growing concerns of the municipalities. Of the many conclusions reached, the Working Party pointed out (Ontario Mineral Aggregate Working Party 1976, p.30) that future demand would increase by as much as 70% by the year 2000. While there appeared to be sufficient resources available to meet this anticipated demand, the Working Party also concluded there is a real danger that without proper resource management, a large part of these resources could be made unavailable through unconcerned municipal landuse planning. The Working Party further concluded that each municipality should be required to make available its "fair share" from the resources under its control, while recognizing the variations in resource endowment from one municipality to the next. The Working Party recommended that the province urge the municipalities "to amend their official plans to incorporate a resource policy to protect [aggregate] resources from conflicting uses." (Ontario Mineral Aggregate Working Party 1976, p.40). So that the municipalities could designate aggregate extraction areas in their official plans, the Aggregate Working Party (Ontario Mineral Aggregate Working Party 1976, p.38) recommended "that the Provincial Government provide all designated regional or county municipalities with the basic surficial geological information on the location and extent of potential mineral aggregate deposits (both stone and sand and gravel)."

A program of providing aggregate resource information to the municipalities in the Toronto area had been initiated in an informal way by the late D.F. Hewitt in the 1960s. However, following publication of the Working Party's report, this program was formalized and expanded under the direction of Owen L. White, Chief, Engineering Terrain Geology Section, who introduced the publication of a series of Aggregate Resources Inventory Papers. At the time of writing, 125 of these papers had been published, covering most of the organized municipalities in southern Ontario and also the more populated areas of northern Ontario.

Engineering Geology Terrain Studies

In its report, the Aggregate Working Party considered the availability of resources in northern Ontario (Ontario Mineral Aggregate Working Party 1976, p.27). The Working Party noted that the available sand and gravel resources accessible by rail will not exceed 500 million imperial tons and that "potentially available resources now identified represent as little as 4% and no more than 12% of the total resources in Southern Ontario", and concluded the "logic of identifying and protecting as many of those areas containing potentially available reserves as possible for future use seems inescapable." The Working Party went on to recommend: "That the Ministry of Natural Resources complete surveys of available resources in northern Ontario and publish data along with a Provincial compilation of aggregate resource data as soon as possible."

While it was possible to conveniently carry out resource assessments for the larger and more populated communities in northern Ontario, the vastness of the area and the great distances between many of the communities made this approach impractical within any reasonable timeframe. The Ontario Geological Survey, in 1977, engaged a consortium of Gartner Lee Associates, Limited;

Figure 2.22. A helicopter involved in the uranium reconnaissance program is used to perform an airborne radiometric survey (1978). This program was jointly funded by the Ontario Geological Survey and the Geological Survey of Canada, from 1975 to 1978, in the search for uranium deposits as an alternative energy source (photograph *courtesy of* D. Wadge, Geophysics/Geochemistry Section).

J.D. Mollard and Associates, Limited; Geo-Analysis, Limited; and Lee Geo-Indicators Limited on the advice of Owen L. White, Chief, Engineering and Terrain Geology Section, and Sherry Yundt, Chief, Industrial Minerals Section, Mineral Resources Branch, and with funding provided by the Ministry of Northern Affairs. The consortium was to carry out "an evaluation of near-surface geological conditions for the purpose of determining the engineering significance of the terrain" (Gartner, Mollard and Roed 1981, p.3), and to illustrate these conditions on 1:100 000 scale maps. These maps would not only show the aggregate resource potential of the Quaternary deposits, their prime purpose, but would also yield information on the engineering characteristics of the surficial deposits useful for transportation-route selection, landfill site location, drift sampling in mineral exploration, construction and other activities (Gartner, Mollard and Roed 1981, p.37–41). The study, under the name "Northern Ontario Engineering Geology Terrain Study" (NOE GTS), took place in 3 stages: in 1977–78, an area of 150 000 km; in 1978–79, 147 000 km; in 1979–80, 73 000 km. By the time the study was completed, the consortium had reported on 103 NTS map units (Gartner, Mollard and Roed 1981, p.6–11).

With funding provided by the Ministry of Natural Resources, the engineering geology terrain study was extended to the Algonquin Region of Southern Ontario in 1979–80, when an additional 30 000 km were mapped (Gartner, Mollard and Roed 1981, p.4). This work, under the name "Southern Ontario Engineering Geology Terrain Study" (SOEGTS) was further extended in the early 1980s to complete coverage of the whole of southeastern Ontario (*see* Figure 2.15d).

Manitoulin Island Study

On the initiative of Peter Telford of the Engineering and Terrain Geology Section, the Ministry of Natural Resources in 1978 provided a grant of $400 000 under the Rural Development Program, and entered into an agreement with the Federal Department of Regional and Economic Expansion to carry out a 3-year study of the limestone/dolostone resources of Manitoulin Island. This project ultimately involved field observations and sampling at more than 3300 sites and 29 boreholes drilled at strategic locations. This work provided a 3-dimensional picture of the stratigraphy and the chemical and engineering properties of the sedimentary rocks, and resulted in the publication of rock-quality maps from which interested parties could select for exploration, areas likely to contain materials with the properties needed to match their particular requirements. The project was successful in encouraging Upper Lakes Shipping and Seeley and Arnill to invest $8 million in a quarry at Meldrum Bay at the west end of the island. This quarry is the only Canadian operation on the Great Lakes producing metallurgical-grade dolostone.

Ontario Geoscience Research Grants Program

In 1976, the Ontario Geological Survey was approached by a small delegation from the University of Toronto concerned with the pending termination of a 5-year "Negotiated Development Grant" from the National Research Council and seeking an alternative source of funding for geoscience research at Ontario universities. For some time, concern had been expressed by exploration geologists that the emphasis of university research leaned too heavily on the fundamental. With this in mind, the writer, together with D.W. Strangway, then chairman of the Geology Department at the University of Toronto, prepared a Cabinet submission requesting permission to set up a research grants program with an annual budget of $500 000 to support applied research at the universities to supplement, but not duplicate in any way the type of work carried out by the Survey. This submission, supported by CUDGO (Chairmen of the University Departments of Geology in Ontario), the Ontario Mining Association, and the Toronto Branch of the CIMM was well received. The Ontario Research Grants Program was put in place in 1977, starting a program that was soon to be recognized in government circles as a model upon which to base further grant programs in other areas of scientific endeavour.

A special feature of the grants program is an annual research seminar at which both grant recipients and the Survey's geologists are expected to present the results of their work to the earth science community at large. The first seminar attracted about 500 delegates, mostly from Canadian exploration companies. It proved particularly valuable, not only by making public the results of the latest research at the universities, but in bringing together geoscientists from academe, government and industry for exchange of information and ideas pertaining to the geology of the province to mutual advantage and to enhance a greater appreciation of each other's roles in the advancement of the science. The seminars, now leaning heavily to "open houses", in which the Survey itself has assumed a major role, have been held annually since 1978. During this time, attendance has increased steadily; and in the late 1980s the seminars have attracted up to 1000 delegates, thus rivalling some of those of the professional societies on the National scene.

Airborne Magnetic-Electromagnetic Surveys

Despite the positive results of airborne magnetic-electromagnetic (AM-AEM) surveys in Quebec in the 1960s, such surveys, as a means of encouraging mineral exploration were resisted by the Ontario Department of Mines on the grounds:

1. They would put the government in the position of conducting exploration for diamond-drill targets, a role which was clearly understood to be that of private enterprise.

2. They would tend to preempt private-sector exploration planning.
3. They could very well duplicate surveys already completed or in the course of being carried out by mining companies, and thus would be received with little enthusiasm, if not outright annoyance.

In 1973, however, the Ontario Geological Survey conducted a questionnaire to determine to what extent this may actually be the case. The questionnaire indicated clearly that industry officials were equally divided on the question as to whether or not the provincial government should sponsor AM-AEM surveys, and that this difference of opinion was just as true for the large exploration company as the small one with most to gain. On the strength of this, and because up to this juncture the policy of nonintervention had precluded the low- income prospector from being able to enjoy the benefits of AM-AEM surveys, it was decided, on the urging of Roger Barlow, Chief, Geophysics/Geochemistry Section, to begin such surveys in selected areas to encourage exploration. The first airborne magnetic-electromagnetic survey was conducted in the Matachewan area in 1975. Since then, such surveys have become a regular feature of the work of the Ontario Geological Survey, particularly in its "community-based" programs (*see* Figure 2.15a).

Community-Based Programs

With the budget and staffing constraints of the late 1970s, the "core program" of the Survey began to suffer in regard to the number of mapping parties placed annually in the field, and the number of geological assistants recruited from the universities to staff those parties. Alternate sources of funding outside the Ministry of Natural Resources became necessary to maintain the work of the Survey at anything like the level prior to 1972. Fortunately, in 1977, the Ministry of Northern Affairs approached the Survey with a recommendation to mount a geological program in the Atikokan area, where the closure of the Steep Rock and Caland iron mines had precipitated economic hardship. The Atikokan Geological Survey (AGS) Program, to include an airborne magnetic-electromagnetic survey of the volcanic rock part of the area, bedrock and Quaternary geological mapping, and mineral deposit studies, was initiated in 1978. Also at about this time, the Economic Development Commissioner of Kirkland Lake, faced with similar economic problems, recommended a major geological program for his area. This program, which came to be known as the Kirkland Lake Initiatives Program (KLIP), was to be similar to the AGS program, but would also include regional basal till sampling. It was funded jointly by the Ontario Ministry of Northern Affairs and the Federal Department of Regional and Economic Expansion (DREE) under the Community and Rural Resource Development Subsidiary Agreement. The Kirkland Lake Initiatives Program was initiated in early 1979 with an airborne geophysical survey. The success of this program, in encouraging mineral exploration and revitalizing the economy of the area, became instrumental in the following years in encouraging the launching of other multidisciplinary programs in northern Ontario, as well as one in southern Ontario, the Southeastern Geological Survey (SOGS) Program, in the latter case funded by DREE and the Ontario Ministry of Natural Resources.

DEVELOPMENTS IN THE 1980s

Ontario Bicentennial Project

Since the first discovery of base metals in the Sudbury area, numerous reports and scientific papers have been written describing and interpreting the geology of the area. However, prior to the 1980s, no comprehensive and multidisciplinary treatise had been published. In early 1981, recognizing this, D.H. Rousell of Laurentian University, suggested to the author that the Ontario Geological Survey call upon the current workers in the area to get together and collectively produce such a memoir, emphasizing this would be a major addition to the scientific literature on the Sudbury Structure. At about the same time, the Ontario Ministry of Culture and Recreation invited other government agencies to consider making contributions to celebrate the bicentenary of the province in 1984. These 2 events and the upcoming centenary of the discovery of nickel-copper ore in the area in 1983, and the fact that 2 of the Survey's geologists, B.O. Dressler and T.L. Muir, were both reaching the final stages of multi-year investigations of the structure, made it appropriate to recommend that the Survey proceed with the proposal. With the enthusiastic co-operation of Falconbridge Limited and Inco Limited, a series of meetings and field trips were organized so that the work in producing the proposed treatise could be closely co-ordinated between geoscientists from government, the 2 companies, the Universities and the Jack Satterly Geochronology Laboratory. The final treatise, accompanied by new coloured geological maps of the Sudbury area, a Bouguer Anomaly gravity map and mine plans of the Onaping–Levack area was published as Ontario Geological Survey Special Volume 1 (Pye et al. 1984) in late 1984. The volume is probably unique in that it gives equal weight to the 2 opposing theories of origin of the Sudbury Structure, the exogenic (meteorite impact) and endogenic (volcanic-tectonic) theories currently being debated, and leaves the readers to reach their own conclusions on the basis of the evidence presented. The volume has been well received and has become the first of a series, of which the present volume on the *Geology of Ontario* is the fourth.

Community-Based Programs—COMDA

The Kirkland Lake Initiatives Program continued until 1983, when it was superseded by a similar program in the Black River–Matheson (BRiM) area along the Porcupine–Destor "break" to the north. The Southeastern Ontario Geological Survey (SOGS) Program ended in 1984. Still another multidisciplinary program had been started in the Opapimiskan Lake–Pickle Lake area where Dome Exploration Limited had discovered a potentially economic gold deposit. The several interministerial and intergovernmental agreements involved in these programs soon proved unwieldy. Then, in 1985, the Ontario and federal governments entered into a $30 million new "shared-cost" agreement, the Canadian–Ontario Mineral Development Agreement (COMDA), a subsidiary of the Economic and Regional Development Agreement of November, 1984. COMDA would replace separate contracts for programs in individual community-based areas. Under its broad umbrella, it would provide $20.8 million for multidisciplinary geoscience programs in: 1) southern Ontario, where the emphasis would be directed toward investigating the potential for industrial minerals and rocks; 2) the Ignace–Sturgeon Lake area of northwestern Ontario, noted for base metal deposits discovered in the 1970s, and also for earlier-discovered and developed gold deposits; 3) the Sudbury–Cobalt area; 4) the Beardmore–Geraldton gold area; 5) the Kenora–Fort Frances area, where there have been discoveries of base metal and gold deposits, and industrial minerals; and 6) the Timmins–Black River–Matheson area, replacing and extending the Black River–Matheson (BRiM) program.

Drill Core Storage Program

The possibility of establishing facilities to store diamond-drill core was considered by the former Department of Mines and Northern Affairs prior to the 1972 reorganization of the provincial government, and the creation of the Ministry of Natural Resources, but was rejected on the basis of the large amount of drill core which was accumulating (approximately 500 000 feet annually) and the cost of a warehouse large enough to store such a huge volume of core.

In the early 1970s, J.A. Robertson, then Chief of the Mineral Deposits Section, brought to the writer's attention that the drill core from many deep drill holes in the Elliot Lake area, were in danger of being discarded by the companies, and indeed a large amount of core already lay in an unprotected state at Massey; and he emphasized that such drill core provided the best stratigraphic sections for future reference and study. At about the same time, following the discovery of the Kidd Creek orebody at Timmins in 1964, a large amount of exploratory drilling had been done in the drift-covered area north of Timmins, where drill core in many instances provided the only hard data, but again was in danger of being lost. Following a report submitted by a committee chaired by R.C. Beard (Resident Geologist, Kenora), the Ontario Board of Industrial Leadership and Development (BILD) agreed to fund the construction and operating costs of several drill core storage facilities in northern Ontario at a cost not to exceed $5 million over a 5-year period.

The first facility was constructed at Swastika and was officially opened June 18, 1984, and the Timmins and Sault Ste. Marie core libraries followed shortly. Additional facilities were also established in 1984 at Thunder Bay in northern Ontario, and at Bancroft and Tweed in southern Ontario. The latest storage facility was officially opened in Kenora on June 20, 1986. Industry response to the Core Library Program generally has been favourable, with greater voluntary donations than originally considered in the design of warehouse capacities.

Exploration Technology Development Program

A decline in the level of mineral exploration in Canada, prior to the resurgence following the major gold discoveries at Hemlo and the federal government's "flow-through share" exploration financing arrangement, seriously affected the largely Ontario-based geophysical service industry, which had won for itself world recognition of its equipment and successful methodologies. When the Ontario Geoscience Research Grants Program was proposed, it was clearly stated the research that would be funded at the universities would result in innovative concepts that could lead to the development of new and better services by the private sector, and that if this were the case, additional funding would be requested for exploration technology development by the private sector. In 1981, the Ontario Board of Industrial Leadership and Development agreed to fund a development assistance program at $1 million dollars annually for a period of 5 years for this specific purpose. The program unfortunately was cancelled in 1986 after the Board was wound up following the change of government in 1985.

Hydrocarbon Energy Resources Program (HERP)

Another 5-year geoscience program funded by the Board of Industrial Leadership and Development was the Hydrocarbon Energy Resources Program. This program, starting in 1981, under the leadership of Peter Telford, was initiated in response to a government decision to attempt to raise provincial energy self-sufficiency to above the 23% level existing in 1979, a decision prompted by upward-spiralling oil prices and insecurity of supply following the actions of OPEC earlier in the decade. The program, funded at $10 million over a 5-year period, consisted of 4 components: 1) an assessment of the potential for economic peat deposits in the province; 2) an assessment of the lignite potential of the Moose River Basin in the James Bay Lowlands; 3) an assessment of the oil shale potential in southern Ontario, where, at Craigleith, oil had been produced from shale in the mid-1860s; and 4) an evaluation of the oil and gas resources of southwestern Ontario, where over 50 000 wells have been drilled and over 2000 are in production. The peatland inventory was concentrated in areas chosen on the basis of development potential and proximity to markets. Sixteen areas, ranging in size from 3000 to 24 000 km², and totalling about 213 000 km², were selected by remote sensing and surveyed and sampled in the field. The work in the Moose River Basin was an extension of a program of geological and geophysical investigations begun in 1975 by the Ontario Geological Survey. It consisted mainly of 2 reconnaissance drilling projects in areas away from the known deposits at Onakawana in the winters of 1983 and 1984. The potentially petroliferous shales of southern Ontario were tested by 60 drill holes and limited geophysical borehole logging. The assessment of the oil and gas resources and resource potential of southwestern Ontario was carried out by the Petroleum Resources Laboratory of the Ministry of Natural Resources at London.

Understanding the Archean

Detailed studies by the geologists of the Ontario Geological Survey and the Geological Survey of Canada, supplemented by numerous precise age determinations by the staff of the Jack Satterly Geochronology Laboratory, have done much in the 1980s to elucidate the makeup and geological history of the Superior Province of the Canadian Shield in Ontario. The subdivision of the Superior Province into distinct volcanic-plutonic, metasedimentary and plutonic subprovinces has been recognized for many years. However, it has been only recently that the composition of these subprovinces, their relationships to one another, and the processes by which they have come into juxtaposition have been satisfactorily explained by calling on modern plate tectonic or mobilist concepts. As early as 1949, Wilson (1949) postulated that the North American continent grew outward by accretion from an early Precambrian nucleus, and in 1968, Goodwin (1968) suggested, on the basis of limited evidence, that the Archean volcanic-plutonic belts in the Superior Province developed successively from north to south. Geochronological studies in the 1980s have now confirmed this suggestion.

Volcanic-plutonic subprovinces such as the Wawa, Abitibi, Wabigoon, Uchi and Sachigo subprovinces are believed to have developed as linear belts of volcanoes that were separated by intervening linear sedimentary basins represented by the Pontiac, Quetico and English River subprovinces. In late Archean time, the volcanic and sedimentary belts were sequentially welded together from north to south to form the Superior Province as we presently know it.

A great many questions about the nature and geological history of the Superior Province remain to be answered, but in recent years, in light of the plate tectonic mobilist theories, considerable advances have been made. Much of the credit for this rests with T.E. Krogh, who greatly improved the U-Pb techniques of precisely dating zircons. The 350 or more dates provided by the Jack Satterly Geochronology Laboratory since its beginning in 1975 have raised the above scenario from one of mere speculation to one of real credibility.

The Reorganization of 1985

The major political event which had far-reaching effects on the Mineral Resources Group was the election of 1985. The Conservative government of William Davis was defeated at the polls and the Liberals, under David Peterson, came to power in Ontario. The Liberals prior to the election had advocated a new Ministry of Mines in response to province-wide criticism of the Ministry of Natural Resources in its handling of mineral management. This was now realized in yet another reorganization, in which the Mineral Resources Group, including the Ontario Geological Survey, was re-united with Northern Affairs to form the present Ministry of Northern Development and Mines (Figure 2.23).

The reorganization of 1985 saw few important changes in the structure of the Survey, except that the Mineral Deposits Section was absorbed by the Precambrian Geology Section under A.C. (Sandy) Colvine, and the Assessment Files Office was attached to the Mines Library under Nancy Thurston.

Greater changes took place in the "field" with the creation of 3 branches or regions, the appointment of John Wood as Regional Director, Northwestern Ontario; the appointment of J.A. (Sandy) McIntosh as Regional Director, Northeastern Ontario; and of A.E. Pitts as Regional Manager, Southern Ontario. In the Northwestern Region the Resident Geologists and the core libraries came under the direction of Mineral Resource Manager K.G. Fenwick; P.E. Giblin held the equivalent position in the Northeastern Region. In Southern Ontario, Resident Geologists and core libraries were placed under the direction of the Regional Manager. At the same time, new Resident Geologists were appointed for the Beardmore–Geraldton, the Schreiber–Hemlo, and Wawa areas in northern Ontario, and for a Southwestern District in southern Ontario, bringing the total number of Resident Geologists in the province to 15. The Resident Geologist's office at Richmond Hill was closed and its responsibilities transferred to the Southwestern District of the Ministry of Natural Resources.

THE OGS AND MINERAL DEVELOPMENT

Miller's contribution to the development of the Cobalt silver camp and the nearby gold camps has been noted, as has Coleman's mapping to the development of the Sudbury nickel-copper field. In the early 1930s, J.E. Thomson mapped the Manitouwadge area and recorded an occurrence of sulphides that 2 decades later was staked and developed as the Geco copper-zinc-silver mine. In the late 1940s, an airborne magnetometer survey resulted in the discovery of the Marmoraton iron mine. These were all important events in the history of mining in Ontario, and provide spectacular examples of the value of the Survey's geological investigations. A policy of the Survey from its beginning has been not to prospect for valuable mineral deposits in their own right, but to leave this function to the private sector. The Survey is not a regulatory body. Its role has been mainly the provision of a publically available scientific data base, through geological mapping and mineral deposit inventory, directed toward locating favourable areas in which others might, through diligent effort, discover valuable resources, and also directed toward devising conceptual models useful in the study of that data base. This is in the tradition of Coleman's work in elucidating the overall form of the Sudbury Structure and the relationship of the nickel-copper deposits to it. It is in this context that the Survey has been very successful in facilitating and encouraging private sector exploration in the province. Occasionally, the Survey's maps have been useful in spotting drill holes. Generally, however, the maps are used for planning exploration programs, for the staking and marketing of mining claims, and as an aid in interpreting private-sector geophysical and geochemical surveys. This is widely recognized by the mineral industry; it is only the naive and inexperienced who would prospect for minerals without the maps, reports, and advice of the

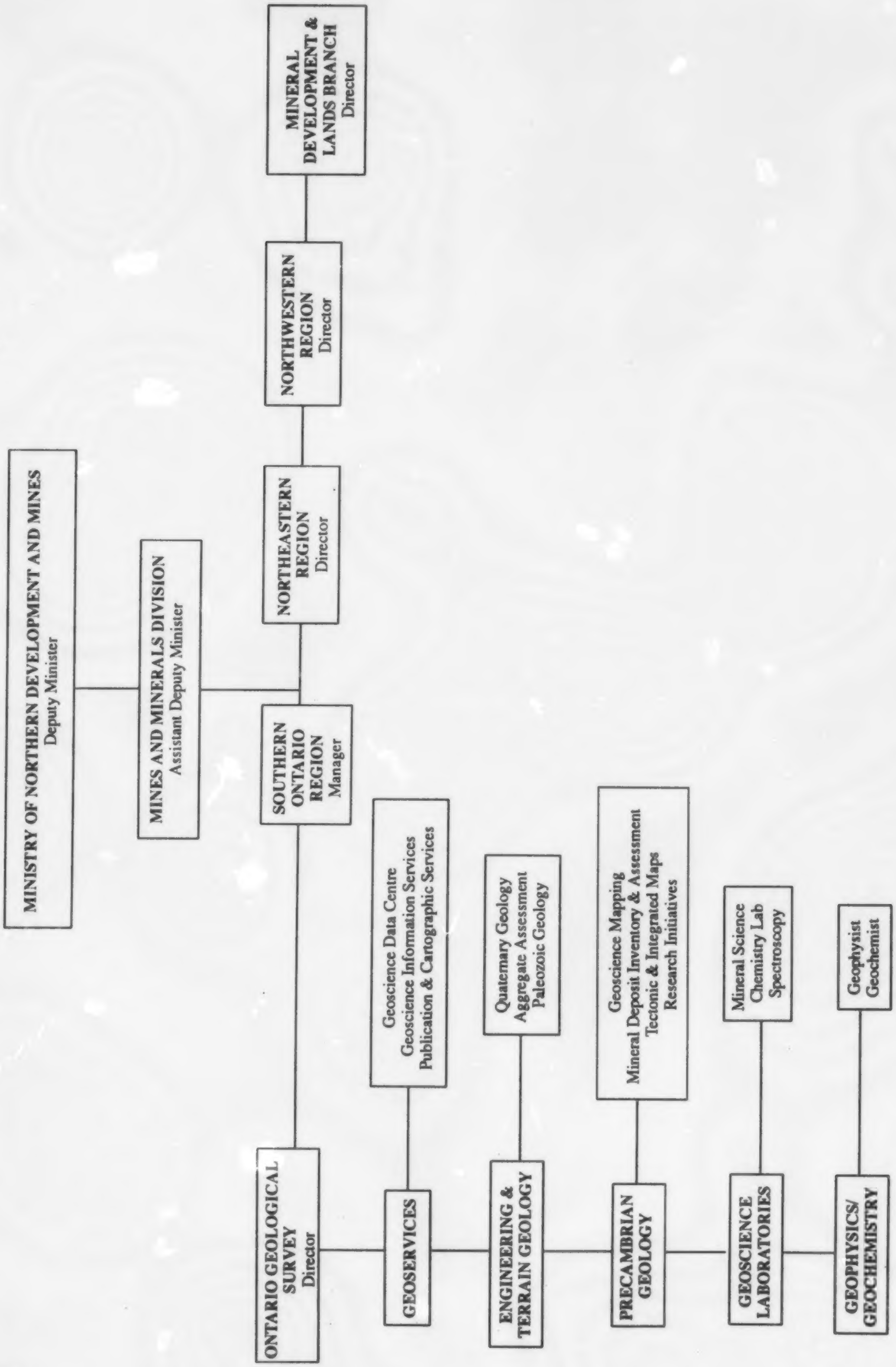
MINISTRY OF NORTHERN DEVELOPMENT AND MINES
Deputy Minister
MINES AND MINERALS DIVISION
Assistant Deputy Minister
ONTARIO GEOLOGICAL SURVEY
Director
SOUTHERN ONTARIO REGION
Manager
NORTHEASTERN REGION
Director
NORTHWESTERN REGION
Director
MINERAL DEVELOPMENT & LANDS BRANCH
Director
GEOSERVICES
Geoscience Data Centre
Geoscience Information Services
Publication & Cartographic Services
ENGINEERING & TERRAIN GEOLOGY
Quaternary Geology
Aggregate Assessment
Paleozoic Geology
PRECAMBRIAN GEOLOGY
Geoscience Mapping
Mineral Deposit Inventory & Assessment
Tectonic & Integrated Maps
Research Initiatives
GEOSCIENCE LABORATORIES
Mineral Science
Chemistry Lab
Spectroscopy
GEOPHYSICS/ GEOCHEMISTRY
Geophysist
Geochemist

Figure 2.23. An organization chart showing the subdivisions of the Ontario Geological Survey and its place within the Ministry of Northern Development and Mines. The arrangement, as shown, was implemented with the reorganization of 1985.

Ministry's geologists. As once said to me by a senior mining company official "the Survey is an important part of the productive sector of the economy and not just another government bureaucracy".

The Ontario Geological Survey and the Ministry's Resident Geologists also play an important indirect role in the development of mineral resources by serving in an advisory capacity to other government agencies on such important issues as road and power line location, land-use and parks planning that might affect exploration and mining, municipal zoning, resource management policies and other matters (Thomson 1970b, p.30).

The Survey has also played an important role in the field of education, a role that likewise has had an impact on the mineral development of the province. As pointed out by Thomson (1970b, p.30–31):

> The importance of practical training to students through government geological field work should not be overlooked. ...University students are employed and trained each summer by the [Ministry]. Many senior students obtain research and thesis material for laboratory studies at universities in order to obtain an advanced degree. Many geologists and officials of Canadian mining companies obtained some of their basic training on Departmental field parties. The influence of Prospectors' and Mineral Exploration Classes since the turn of the...century...must have led indirectly, and possibly unwittingly, to the discovery and development of many mineral deposits. Also, talks to students on career possibilities in geology and the mineral industry have influenced individuals to make this a chosen profession.

THE FUTURE

On July 30, 1986, David Peterson, Premier of Ontario announced that it was the new Liberal government's intention to relocate the Ministry of Northern Development and Mines, including the Ontario Geological Survey, to Sudbury. This relocation is scheduled for 1992 and will shortly follow the centennial of the founding of the Ontario Bureau of Mines. The relocation of the Ontario Geological Survey to the Laurentian University campus in Sudbury will mark the beginning of a new era for the Survey (Figure 2.24).

The most important function of the Survey is geological mapping. Since the early 1960s, geological mapping of the Province has proceeded rapidly. There are now few areas for which geological maps and reports are not available. However, geological mapping is a never-ending requirement because all maps must be updated periodically if they are to be useful tools for mineral exploration. While the rocks do not change, the information to be gleaned from them increases with every advance of the science. This has been discussed at length by Pettijohn (1984, p.244–249), whose text is worthy of the attention of any geologist-administrator faced with the problem of convincing his superiors and funding agencies of the basic necessity of periodically re-investigating and re-mapping even the most highly studied areas.

Over the past century, the Ontario Geological Survey has made a major contribution to the mineral industry and economy of the province, partly by drawing the prospectors' attention to areas with a geological framework and geological history favourable for the occurrence of economic mineral deposits, partly by research and the development of conceptual models of ore genesis, and occasionally by the actual discovery of valuable deposits. This contribution is expected to continue.

Geological surveys, together with geophysical and geochemical surveys, will remain the principal work of the Ontario Geological Survey for the foreseeable future. But this work will be increasingly supplemented by applied research

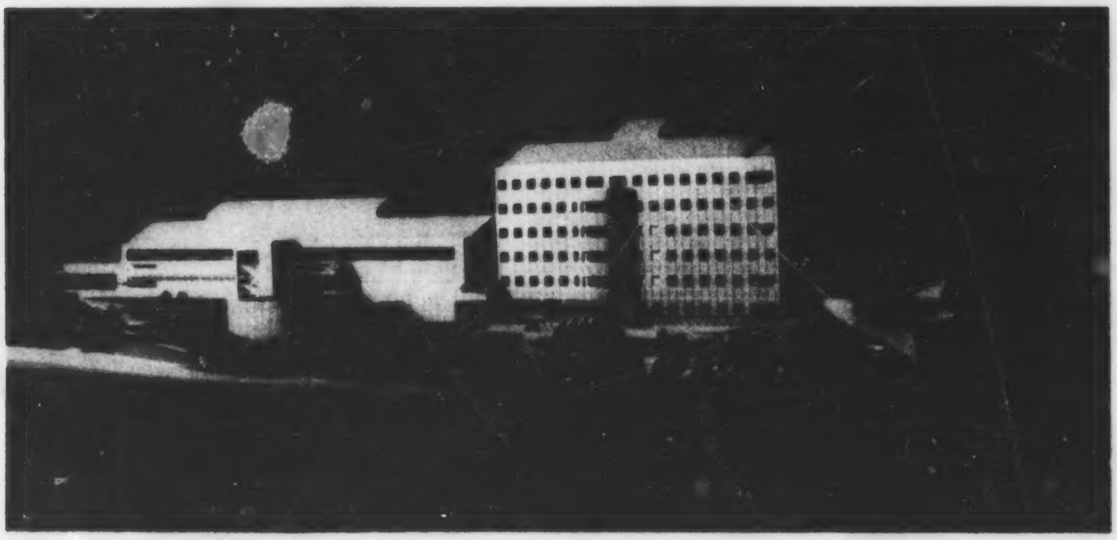

Figure 2.24. A model of the new facilities under construction on the Laurentian University campus in Sudbury. This building will be the headquarters for the Ontario Geological Survey in 1992. The move is part of the Ontario government's program of decentralization of government functions (photograph *courtesy of* Ministry Relocation Project, Ministry of Northern Development and Mines).

as mineral exploration itself changes toward the search for more deeply buried, and less easily detected ore deposits. Geochronology and geochemistry, and even diamond drilling, will become increasingly common as investigative tools. The OGS will become more heavily involved in environmental geology as waste management problems become progressively more acute. Also, no doubt, construction activities will lead to heavier involvement in engineering geology. Politically motivated, community-based programs will become the norm. Computer applications will become prevalent in many areas of the science. However, it will be largely the initiative of the Survey itself which will lead it into new areas of responsibility. The future, despite the setbacks which will occur from time to time, is believed to be a promising one that will provide many career opportunities.

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Chapter 3

The History and Economic Role of Ontario's Mineral Sector

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Abstract

This chapter provides an overview of the role of mining in Ontario's history and economy. The geological and historical heritage of Ontario's northern region is such that the development of mineral resources has the greatest potential to become a significant input into the industrial production process.

The sequence of major discoveries in Ontario began with the construction of railways in the mid-1800s and was followed by: the development of major precious metal mining camps in the early 1900s; and major base metal and uranium camps in the mid-1900s. Since 1980, there has been sustained interest in gold exploration and development. The successive stages of the mining sequence, (exploration-development-mining-processing-sales and profits), are preceded by demands for commodities, consumer goods and supply of investment capital.

In terms of economic impact, mining in Ontario compares very favourably with other sectors, and Ontario has an important status in world mineral production. The challenges and opportunities facing the industry through the 1990s and beyond are: declining real metal prices; a declining land base open to mineral exploration; and increasing institutional constraints such as the legal and regulatory climate in which the industry operates. These constraints are manifested as increased mandated costs. The responses to meet these challenges must include increasing productivity and product diversification through research and development.

INTRODUCTION

Mining and Economic Development

The role of Ontario's mineral sector is very significant when compared with other sectors of Ontario's economy. This chapter provides an overview of the role of mining in Ontario's history and economy. A broad overview of the general relationships between mining and economic development provides a context for, and gives meaning to, the manifold aspects of the rise and role of Ontario's mineral industry.

For this overview of the relationships between mining and economic development, geological and historical factors are simplified as follows. Based on geology, the earth is divided into 2 basic types of surfaces. Early centres of civilization rose in areas of geologically young overburden and temperate climate. This favoured the development of agriculture and of relatively high population densities. The latter made a high degree of division of labour possible. The heavy overburden, however, would have required highly capital intensive methods, for mineral exploration and extraction and these are methods which require prior development of an entrepreneurial class. The "hydraulic civilizations" (Quigley 1961), characteristic of places such as the Nile Valley and Mesopotamia, inhibited the development of such a class. One can ignore the few mining operations of the hydraulic civilizations which were state-run and slave-operated, as were the great engineering undertakings which made such civilizations famous.

These civilizations had to get their metals from the border zones of the other major geological regime. These were initially areas of low population density and often intemperate climate in the mountainous areas of Western Europe and the Near East, and, subsequently, in the mountains and Precambrian shields of the American West and northern Canada, Siberia, parts of Australia, Africa and South America. In these areas, metalliferous formations and structures were widely exposed. Such surfaces provided poor foundations for agriculture, but allowed for easy visual discovery of outcrops of mineral occurrences by prospectors with only crude technical aids, and also allowed for fairly easy extraction.

In Europe, these physical factors were complemented and rendered highly productive by a combination of cultural factors which nourished the roots of crucial institutional developments. Nobel Laureate, F.A. Hayek (1973, 1988), pointed out that the extended traditional order of Europe was essentially an evolved process encompassing morals and law, governments and markets. This order made it possible to sustain more from discovered resources and to discover more resources than was possible under the command systems that had been instituted everywhere else to facilitate the achievement of specific results.

Through the Middle Ages and the Industrial Revolution, such developments allowed European nations to spread the application of capital and entrepreneurship to the mineral resources of "less-developed" areas where previously isolated communities survived under tribal organization which did not encourage capital acquisition or the development of entrepreneurship. Thus, for centuries, geology may be seen as having determined the spread of civilization via mineral development through many areas that share crucial characteristics with Ontario's north.

Geological and historical heritage, manifested in low population density, lack of agricultural base, long distances from population centres and inclement weather still affect the fortunes of Ontario's northern regions. Our mineral potential is the only major future source of wealth that can, on a reasonably large industrial scale, be developed in northern Ontario to form major inputs into industrial production processes.

THE HISTORY

The Sequence of Major Events[1]

During their earliest stages, the economies of Ontario and Canada were characterized by the dualism of subsistence farming and the export of goods for which "finding" was tantamount to production. Exported goods included: fur, fish, lumber and precious metals from placers or high-grade veins. As most of southern Ontario is overlain by Mesozoic and Paleozoic sedimentary rocks, mineral discoveries in Ontario on the whole began rather late. Many of the first and some of the major discoveries occurred during railway construction through the Precambrian Shield areas of the province. In 1846, copper deposits were discovered near Bruce Mines on the North Shore of Lake Huron and silver was reported from the vicinity of Thunder Bay.[1] In 1866, the first gold was found near Madoc and in 1883 the copper-nickel ores of Sudbury were discovered (Figure 3.1).

Some iron mining activity in Ontario during the 19th century centred around deposits in the southeast and bog-iron deposits in the southwestern part of the province. These deposits formed the basis for a number of local subsistence smelters and foundries. Later, the Helen, Magpie and Moose Mountain mines, as well as a few smaller ones, produced iron from 1904 to 1926, but then there was no iron ore production for about a decade and a half while the young Ontario steel industry had to import feed from the USA.

In economic terms, the rush or boom phase of a new country's mineral industry development may be explained as follows: as long as both people and capital are scarce, only the discovery of a mineral occurrence that can be turned into a valuable commodity with a small amount of labour and investment will attract attention. The apparent magnitude of the return to labour, once a strike has been reported, attracts large numbers of people and the rush is on. Commonly, many expectations are disappointed and the relative price of labour in the typical early boom camp declines drastically. As a result, capital is attracted and if deposits extend to depth, a conventional mining camp grows, around which a viable community with a more diversified economic base develops. However, if the mineral resource base is too small, the camp may turn into a ghost town. Early Ontario mining has experienced both types of situations.

The first prospecting rushes for precious metals in Ontario occurred near Eldorado in Hastings County in 1866, in the Lake of the Woods area in the late 1870s and early 1890s, and on Silver Islet in Lake Superior in 1868. These rushes did not lead to the development of viable communities. The major precious metal mining camps in Ontario developed after the Sullivan Camp in British Columbia was well established and after the Klondike rush had passed its peak in 1900. In 1903, silver and cobalt were discovered at Long Lake near Cobalt. In 1906, gold was found at Kirkland Lake, in 1908 in the Porcupine area and in 1925 near Red Lake. This was the end of developments that were essentially not influenced by mineral policy.

The 1930s saw 3 actions which were to affect Ontario mining profoundly. By 1934, the US increased the guaranteed gold price to about US $35 per ounce, that is at over 60% higher than it had been since 1820. This gave gold mining developments in Ontario and all of Canada a tremendous boost during the later stages of the depression. After the Second World War, when the US gold price had changed its nature from a floor price to a ceiling price due to increasing cost and ownership constraints, Canadian producers benefited from the Emergency Gold Mining Assistance Act (EGMA), essentially a cost-plus formula under which they

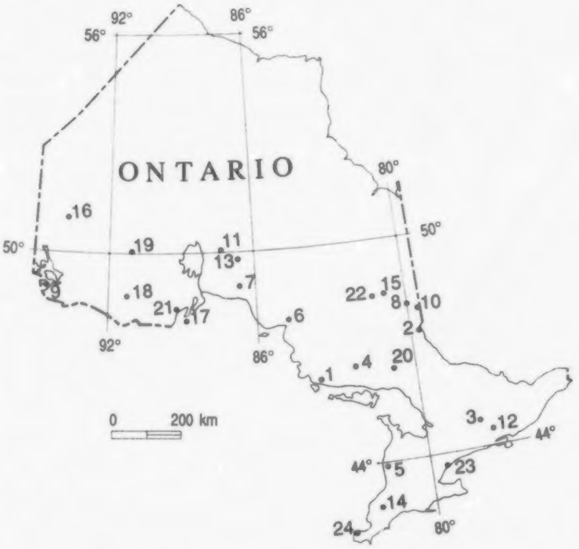

Figure 3.1. Ontario mining locations: 1-Bruce Mines, 2-Cobalt, 3-Eldorado, 4-Elliot Lake, 5-Goderich, 6-Helen Mine, 7-Hemlo, 8-Kirkland Lake, 9-Lake of the Woods, 10-Larder Lake, 11-Little Longlac, 12-Madoc, 13-Manitouwadge, 14-Oil Springs, 15-Porcupine, 16-Red Lake, 17-Silver Islet, 18-Steep Rock Lake, 19-Sturgeon Lake area, 20-Sudbury, 21-Thunder Bay, 22-Timmins, 23-Toronto and 24-Windsor.

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[1] *The historical and statistical data in this and subsequent sections of the chapter have been assembled and updated by the author from many issues of the* Annual Report of the Ontario Department of Mines *and its successor publications;* The Canadian Minerals Yearbook, *Energy, Mines and Resources, Ottawa;* Ontario Statistics 1986, *Ontario Ministry of Treasury and Economics; the* Canadian Mines Handbook, *Northern Miner Press, Toronto; and Green (1983). See also Smith (1986) and Anders et al. (1975).*

could sell their gold production to the federal government. This act, designed to protect Canadian gold mining against total collapse, was necessary until 1972 when rising gold prices made it redundant. Another depression measure intended to benefit the mining industry was a 1936 amendment to the Federal Income Tax Act which exempted new producing metal mines for 3 years. This was a powerful incentive until it was revoked and replaced by other provisions. In 1939, Canadian base metal producers agreed to supply the Canadian government with copper, lead and zinc at fixed prewar prices and the industry was put on a war footing.

Iron ore shipments from Steep Rock Lake commenced in 1944, although the existence of an orebody under the lake had been predicted in 1897 (McInnes 1899), 1902 (Miller 1903) and 1925 (Tanton 1927). In 1953, the zinc-copper deposits at Manitouwadge were discovered. Large-scale uranium production began at Elliot Lake in 1956. The zinc-copper-silver discovery by Texas Gulf Sulphur Co. Inc. in Timmins (the Kidd Creek Mine) occurred in 1964, the most significant base metal discovery in Ontario since Sudbury.

The extraordinarily high gold prices of 1980 and subsequent measures such as the Ontario Mineral Exploration Program (OMEP) and the federal government's "flow-through" shares program, initiated a new period of feverish gold exploration and gold mine development activity in the early 1980s that is still continuing. The most significant result of this was the development of the 3 Hemlo gold mines. However, there were many other new gold properties that came into production, that were reopened, or, that initiated major capital programs to increase capacity. The result was that gold production in Ontario, which had reached a low point in 1981 due to the low gold price in the immediate post-WW II decades, had by 1989 more than quadrupled in volume. This volume, however, is still just under 80% of the highest production level in 1940.

Metal mining presently accounts for 77.3% of the value of all Ontario mineral production and virtually all of this production is from northern Ontario. Of the other subsectors, where production takes place primarily in the southern part of the province, structural materials, sand, gravel and crushed stone are most important. These materials account for 17.7% of mineral production. Oil and gas are concentrated in southwestern Ontario where, in 1858, the first oil well in North America began to produce near Oil Springs, 1 year before the famous oil discovery near Titusville, Pennsylvania. Oil and gas account for 1.2% and industrial minerals account for 3.8% of the value of Ontario mineral production. Over half of the value of nonmetallic or industrial mineral production is accounted for by salt from 2 mines; 1 is south of Windsor, the other mine is at Goderich. The other major products, in descending order of importance, are by-product sulphur produced from smelter gas after base metal refining, nepheline syenite, gypsum and soapstone and talc.

In 1989 and 1990, changes to "flow-through" share financing and termination of federal exploration incentives have, to some extent, been offset by increases in provincial incentives. It may be expected that a reasonable level of exploration and mine development will continue through the 90s, albeit perhaps not at quite the hectic pace of some recent years. Both the historical record and current exploration activity tell us that the national and international investment communities have judged Ontario to be a good place to put exploration and development dollars and to continue to do so.

The Mining Sequence

Over 100 years ago, the French economist and politician, Frederic Bastiat wrote a brilliant essay (Bastiat 1864) on the importance of distinguishing between what is seen and what is not seen, and this is a distinction that we have to keep in mind when looking at the mining sequence.

In the mineral industries there is a visible physical sequence of activities: exploration-development-mining-processing-sales and profits. Judging on the basis of what is seen, it might appear that great changes in the level of activity can be achieved by focussing on the beginning of the sequence, such as by concentrating policy measures in the areas of support or incentives to the exploration sector.

The above sequence, however, is preceded by 2 sequences of economic decisions that cannot be seen. The unseen sequences come together at the point at which the visible sequence begins, as at the fulcrum of a pair of scissors. On the demand side, one starts with consumer demand for final goods, which utilize some mineral products or metals. This leads to a demand for capital goods that incorporate more metallic mineral commodities, and, leads further to a demand for orebodies that are supplied by the exploration industry. On the supply side, the sequence starts with expected profits for investors that lead to investment decisions in favour of mining, rather than in alternatives. These in turn lead to a decision to acquire discovered mineral property, and ultimately, to explore for new orebodies.

The outputs of the mining industry are mineral commodities. Demand for mineral commodities derives from the demand for consumer and capital goods. The primary inputs to the mining industry are orebodies which are the major output of the exploration industry. Demand for orebodies is derived from the demand for mineral commodities. Among the inputs to the exploration industry are geological and geophysical services, diamond drilling, line cutting and others which are the output of the exploration service industries. Demand for these services are derived from the demand for orebodies. Not much can be done to directly stimulate a derived demand. Action is more efficiently directed at the supply side such as reducing the costs of and thus increasing the competitiveness in the delivery of mineral commodities to their primary users worldwide.

Reductions of extraction cost will be reflected in higher rates of return on investment in mining and in an increased demand for orebodies, and therefore exploration activity. Domestic demand for the products of the mining and exploration industries, mineral commodities and orebodies will, for the foreseeable future, remain inadequate to sustain a healthy level of activity and the industry will remain export-dependant.

In recent years, the high level of activity in mineral exploration has been directed mostly at gold. This suggests that it has been supported by speculative elements in the minds of prospectors, developers and mine financiers. Such elements are part of every business or investment decision. A more solid base for the development of downstream industries is more likely to be provided by base metals because of their continuous and large-scale industrial use.

The Pattern of Prospecting Activity

Figure 3.2 shows the pattern of claim-staking activity in Ontario since the beginning of this century. It shows the number of new claims staked annually and indicators of the net changes in total number of claims in good standing. The pattern reflects fundamental changes in the output of Ontario mines, as well as in exploration technology and claim-staking practice. Each staking peak corresponds to a particular major discovery or follows a significant real price change, such as the early 1934 increase in the "floor price" for gold by the USA, the 1947–1974 inflation-induced base metals boom and the 1980 peak in the gold price.

The initial product change was from precious to base metals: this represented a continuous decline in the value to volume ratio of ore mined. Mining activity in any particular sector was characterized by a general shift from higher to lower grade ores with a temporary reversal of this trend in the 60s. There was as well a shift from vein-type to disseminated type deposits.

Whereas, at the beginning of the century prospectors looked for outcrops; deposits are now commonly found under thick overburden. Prospecting moved from labour intensive work to high capital-intensive geophysical and geochemical methods. In the initial stages, these methods are now commonly airborne surveys that cover large tracts of land in order to narrow down target areas for closer investigation. These changes required the control and staking of large areas. This is reflected in the changing pattern of mining claims recorded. Up to the early 40s, "normal" staking activity fluctuated around 5000 claims per year with "gold rush" peaks reaching 15 000 to 20 000. In the post-WWII period, the "normal" activity was between 13 000 and 20 000, with base metal peaks reaching 45 000 to 55 000, and in the early 80s "gold-fever" drove annual claims recorded above 65 000.

To obtain some perspective on the relationship of exploration to mining, one may compare estimates of the costs of finding a new deposit (excluding extensions of known deposits) with the costs of developing an orebody to the production stage. Related to current interest rates, the average cost of finding a mine amounts to about 1 year's interest on the cost of developing it. Comparing actual expenditures in the Canadian metal-mining sector, which have been fluctuating rather widely in recent years, the following picture emerges. If outside exploration expenditure is assigned a weight of 1, then on-property exploration and development expenditure have a weight of 5 to 10, capital and repair expenditures a weight of 40 to 60, direct labour costs a weight of 25 to 30 and government-imposed costs a weight of 30 to 50. The last item includes environmental costs, mandated indirect labour costs, taxation and all compliance costs. Thus, measures to reduce costs of exploration appear as lowering the cost or increasing the efficiency of a factor with a weight of 1, versus lowering the costs or increasing the efficiency of other factors with a combined weight between 100 and 150.

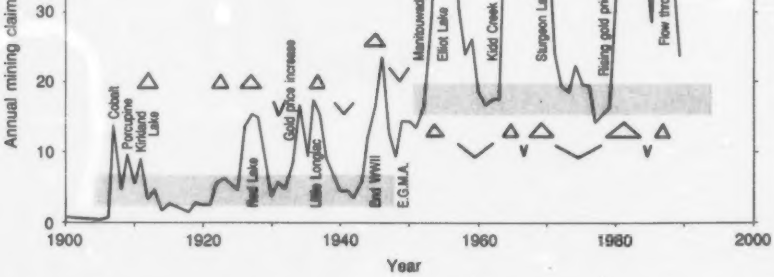


Figure 3.2. The pattern of prospecting activity. The number of mining claims recorded annually is combined with the net change in the total number of claims in good standing to show the change in commodity emphasis and the change in exploration technology and staking practice. E.G.M.A. = Emergency Gold Mining Assistance Act.

Although exploration expenditures only carry a weight of 1, without them none of the others would be incurred.

The Pattern of Mining Activity

Coincident with the changes in the pattern of exploration and prospecting activities, the mining industry itself underwent some fundamental changes. Figure 3.3 shows levels of average annual direct employment in Ontario metal mining and physical output in short tons of ore mined. Both series combine precious and base metals. From the beginning of the century to the end of WWII, both sets of data follow similar upward patterns: declines are associated with the 1920 recession, the 1929–1932 depression and WWII. From 1900 to 1960, annual tonnage increased at an increasing rate, while employment increased at a decreasing rate, particularly in the 1946 to 1960 period. Declines during the 1920–1921 recession and the 1929–1932 depression are followed by spectacular growth periods. The industry recovered in the late 40s and in the 50s from a post-WWII slump. The decline of the early 60s in both employment and physical output ended in a remarkable recovery, as far as output was concerned, while employment trends continued to decline. From the early 70s on, physical output again began to decline in step with employment levels.

Figure 3.4 shows, in constant 1961 Canadian dollars, the trend in real total value of production in metal mining against the trend in real unit value of metallic ores. Average value per ton of Ontario ore increased during the period 1900 to 1911 from about $20 to the $70 range, after 1911, the value decreased dramatically to the end of WWII, a decrease broken during the 30s due to the increase in the guaranteed floor price for gold by the US. Since the mid-40s, the average real value of ore has remained fairly steady at $10 to $15 per short ton, with a peak due to relatively high metal prices (particularly of gold) in 1980 and then returning to the $10 to $15 range. Only in the last year, has there been again a small increase which may be attributed to the almost exclusive emphasis on development of new gold mines.

The initial decline in average value of Ontario ores, in the second and third decade of this century, was a result of the exhaustion of the relatively small high-grade precious metal deposits, the discovery of which had triggered the boom phase of the first decade of the century. Later, decline was associated with the shift from precious to base metal mining. The decline in average value of ore is in contrast to the steady increase in total value of production in real terms from 1900 until the 1960s and 1970s. Until 1960, there were only 2 major declines in total values—1 decline is associated with WWI, the other decline with the WWII period. These declines in total volume correspond to similar decreases in annual average employment, with the WWII period also corresponding to a decline in output of ore mined. The decrease in the real value of production during the early 1960s parallels the simultaneous fall of annual average employment and annual short tons of ore mined; this decrease in value was followed by an increase during the late 1960s. The overall trend in real value of production from 1960 to the present is much flatter than during the preceding periods. Particularly noticeable is the great volatility of production values since the late 1970s.

The divergence of declining average ore values and increasing real values of production through this century can largely be accounted for by changes in technology which are reflected in average mine size.

As the value to volume ratio decreased, the average size of mines increased. Thirty to 40 years ago, gold and base metal mines commonly produced 300 to 1500 tons per day,

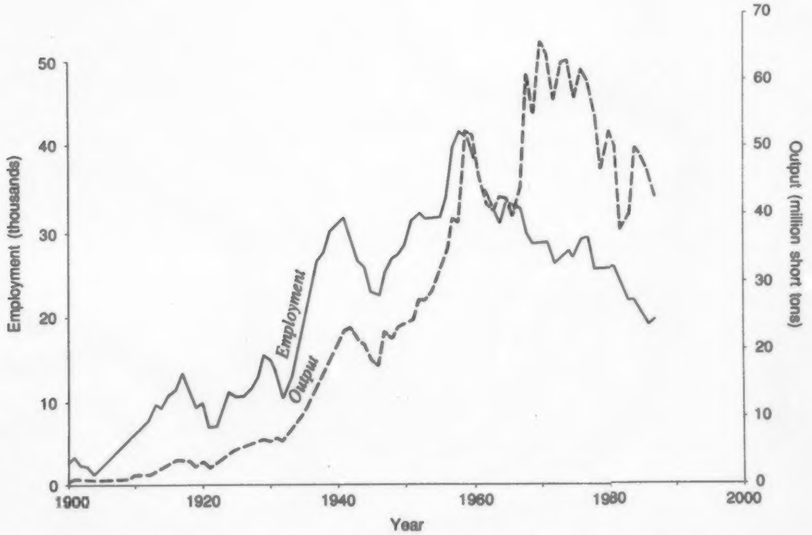


Figure 3.3. Ontario metal mining—annual employment and output. Precious and base metal data are combined. Trends between average annual direct employment and annual physical output in short tons of mined ore are similar.

today the norm is 3000 tons per day and more. The individual miner working with jackleg, scraper or mucking machine, and ore train has been replaced by the drill jumbo and scoop tram operator. The value of a miner's equipment has increased from several hundred or at most a few thousand dollars to tens of thousands or even over a hundred thousand dollars. Mill size and complexity have increased as well.

Average mine size increased from 1900 to 1980 by a factor of about 100 (Figure 3.5). From 1980 to 1987, there has been a decline of 21.5%. The increase, from 1900 to 1980, reflects a general shift from small-scale precious metal mining to large-scale mining of massive or disseminated base metal deposits. The decline since 1980 reflects the increasing role played by new gold mines.

Such changes, which have spelled increasing capital requirements and longer lead times from discovery to production, have ultimately affected mine financing and industry structure. These changes have been reinforced by concurrent changes in the labour market and in the role of government *vis-à-vis* the industry.

We may now attempt to generalize from this record as well as from considerations of general mining statistics and from arguments developed in specialized reports. Developments may be said to have occurred in a logical sequence in

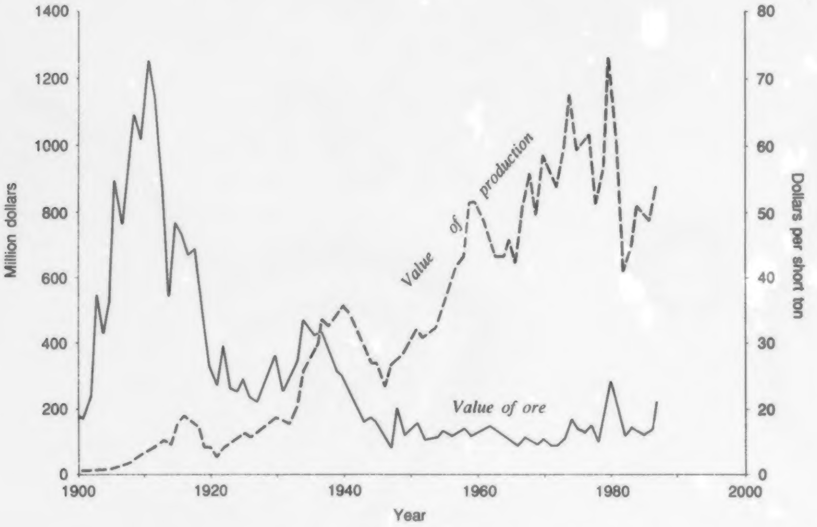


Figure 3.4. Value of production and value of ore. The real total value of production in metal mining is measured in constant 1961 Canadian dollars against the real unit value, in dollars per short ton, of metallic ores.

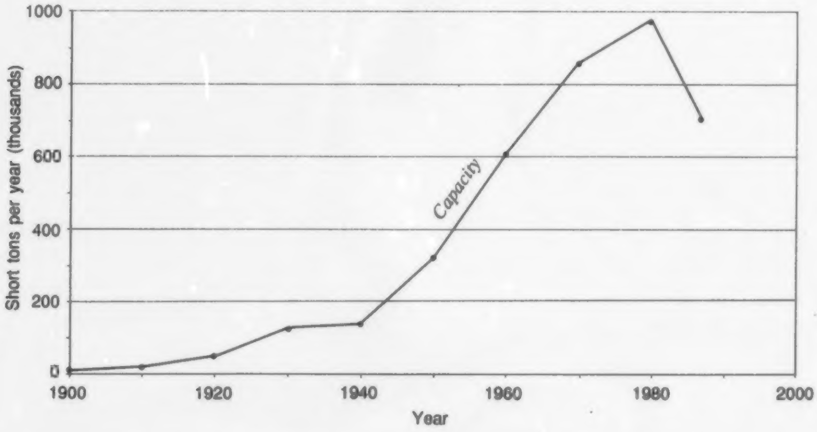


Figure 3.5. Changes in average metal mine capacity. The actual pattern of capacity increase decade by decade in annual short tons of average metal mine production is as follows: 1900–9800; 1910–19 100; 1920–50 900; 1930–130 100; 1940–140 700; 1950–328 500; 1960–613 000; 1970–863 900; 1980–985 700; and 1987–700 700.

closely interconnected areas. The initial change was one in product which led to changes in the type of deposit primarily mined in the province. Different deposits required different modes of operation and of technology in prospecting, development and production. These changes, together with concurrent changes in the nature of government-industry relations, led lastly to changes in dominant modes of financing and in industry structure.

THE ECONOMIC ROLE

Sectoral Comparisons

The economic role of Ontario's mineral sector, in terms of wealth creation, compares favourably with other sectors of Ontario's economy, particularly in northern Ontario. The question remains whether northern Ontario's land and labour is most efficiently employed in mining.

In Ontario, the areas directly affected by mining total approximately 200 km^2 or about 75 square miles (Figure 3.6). During the life of mining operations, these areas are not available for other uses. Areas affected by mining encompass the surface areas required for access to and operation of mines, including tailings dams. The total land area occupied by mining operations (200 km^2) is roughly equal to 5.2% of the total area occupied by roads; or 0.7% of staked, patented and leased lands; or 0.02% of Ontario's total land area.

In recent years, the value productivity of mining is about $25 million/km^{2}pa (*see* Table 3.1a). Agriculture, from a total area of about 60 000 km^2, has an average value productivity of about $85 000/km^{2}pa; figures for particular crops varying from $41 000/km^{2}pa for wheat to $380 000/km^{2}pa for fruit. Agricultural production is primarily from parts of southern Ontario where there is no metal mining activity and is rarely practicable in northern Ontario where mining prevails.

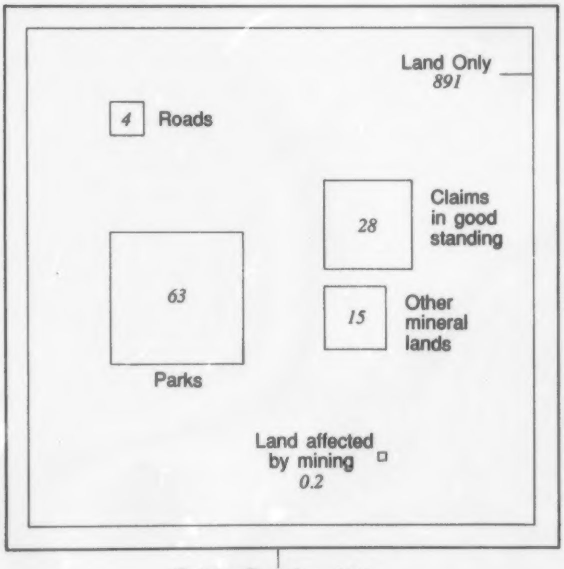


Figure 3.6. Ontario land affected by mining (in thousands of square kilometres). Land "affected" by mining is a generous measure of the total area not concurrently available for other uses. "Other mineral lands" are patented mining lands, exploratory gas and oil licences, gas production leases, exploratory licences of occupation, mining licences of occupation, and mining leases.

The value of "shipments of roundwood" used by the logging industry is suitable for comparison with the value of mineral production, as published in Ontario statistics. On this basis, over the past quarter century, the total output value of the forestry industry has varied between 1/5 and 1/7 that of the value of metallic mineral production. On a productivity-of-land basis, comparison depends largely on the definition of forest land used in forestry production. Using "productive forest land" only, the most favourable definition for the forestry sector, the value productivity would be $1660/km^{2}pa. In terms of total forest land the value would be about $780. This would compare with a value productivity for mining of $116 000 for mining based on the area of "claims in good standing" and "other mineral lands" in addition to "lands affected by mining."

For park lands, reliable data for value productivity are not available. An approximation may be derived by the following: total Ontario expenditures by all travellers, divided by the number of person trips, multiplied by the number of provincial park visitations, divided by the area covered by provincial parks. The resulting value productivity is about $9200/km^{2}pa. Travellers' expenditures are highly biased towards southern Ontario convention business, but it is the only comparison figure available.

A detailed analysis of the opportunity cost of mining in the north was published in: "The Economics of Mineral Extraction" in 1980 (Anders et al. 1980). It concluded that if all areas affected by mining were to be used for agricultural purposes, total agriculture receipts would have increased by between 0.1% and 1%. Total provincial receipts from forestry would have increased by between 0.02% and 1.3%. Both for forestry and agriculture, percentage increases would have been below the limit within which total revenues could be calculated in the first place. If all such disturbed land were returned to hunting, the result was an absurdly tiny amount.

We can express the annual value produced in terms of multiples of the lowest value for forestry. In these terms, the highest forestry value would be 2.1, the value for parks about 11.8, the value for agriculture 109.0, while the value for mining would be 32 000.

An important measure of the relative economic importance of the mining sector is the value added per employee per year; mining leads with about $58 700, which is almost twice as much as the average for manufacturing in Ontario ($32 400), and significantly higher than other primary sectors such as forestry ($29 200) and agriculture ($19 200). In the manufacturing sector, the chemical products have the highest value added per employee per year ($48 100), followed by the food and beverage industries ($37 300), the primary metals industries ($36 800), transportation equipment ($32 800), and the wood working industries ($20 000).

Table 3.1. Productivity comparisons—Ontario metal mining. **a)** Value of land productivity per square kilometre per year. **b)** Value of labour added per employee per year.

	(a) LAND	
Sector	**$/km^{2}pa**	**Multiples**
Mining	25 000 000	32 050
Agriculture		
Average	85 000	109
Wheat	41 000	53
Fruit	380 000	487
Parks	9 200	12
Forestry		
High	1 660	2
Low	780	1
Hunting	50	0.06

	(b) LABOUR	
Sector	**Value added per man–year ($)**	**Wages ($/wk)**
Mining	58 700	691
Manufacturing	32 400	504
Forestry	29 200	566
Agriculture	19 100	

The high value added per worker in mining, compared to other industries, is reflected in comparative wages earned in the mineral sector. In 1986, metal mining leads other industry groups on a national basis, with average weekly earnings of $691. The industrial aggregate being $431, earnings for other industry groups are: forestry $556 and manufacturing $504. All other sectors lagged far behind (*see* Table 3.1b).

In percentage terms, the value-added per employee per year in metal mining is about 90% higher than the average for all workers, 150% higher than for textile workers, 44% higher than in the transportation equipment industry and 38% higher than in the chemical product industries. With regard to earnings, metal mining average weekly wages are 37% higher than the manufacturing average, and 22% higher than forestry. The purchasing power generated by Ontario mineral sector employees, most of it in northern Ontario, must not be neglected as it is a result of total wages and salaries of about $1.2 billion a year. The economic impact of other expenditures must be added to these wages and salaries. Currently, Ontario's mining industry spends over $700 million per year on process supplies, and almost $500 million for fuel and electricity.

Metal mining, therefore, compares favourably with other activities. Differently put, every job transferred from any other industry to metal mining makes us all richer, and every job transferred from metal mining to another industry makes all of us poorer. Moreover, while direct employment in metal mining accounts for less than 1% of Ontario's work force, the value of mineral products exported pays for over 20% of our imports.

International and Productivity Comparisons

Ontario is one-tenth the land area of Canada or twice that of France, or one and a half times that of Texas. If Ontario were an independent nation, it would rank among the world's top ten metallic mineral producers and exporters. Ontario would be the world's second largest nickel producer after the USSR, the fourth largest producer of zinc, the fourth largest producer of cobalt, the seventh largest producer of gold, the seventh largest producer of silver and the eighth largest producer of copper.

The value of metallic mineral commodities produced in Ontario at the primary stage fluctuates between 40 to 50% of the value of metallic mineral commodities produced by the United States. Ontario is Canada's largest producer of metallic minerals, producing about 3 times as much as Quebec or B.C. and about 40% of the nation's total. While direct employment in metal mining accounts for less than 1% of Ontario's work force, the value of mineral products exported pays for over 20% of our imports. These figures fluctuate somewhat depending on relative metal prices and Canadian dollar exchange rates.

This mineral production takes place next door to the United States, the world's largest consumer of metallic mineral commodities and Ontario's best market and most prominent provider of capital. For Canada as a whole, the share of GNP derived from metal mining is almost ten times higher than the United States or about one hundred times higher than West Germany.

Figure 3.7 correlates the value and volume of production per man-year. The 2 sets of data show similar rising trends with both trends steepening significantly in the early 1960s. In general, from the mid-1960s on, production volume rose faster and declined less rapidly than value of production. This is most likely a result of technological change, particularly the increasing application of bulk-mining methods. The flattening of both trends since the late 60s, is associated with the declining rate of increase in average mine size.

CHALLENGES AND OPPORTUNITIES

Declining Real Prices, Declining Land Base and Increasing Institutional Constraints

In the future, Ontario's mineral sector will be faced by 3 major challenges: 1) declining real prices for mineral commodities on world markets; 2) a declining land base open for

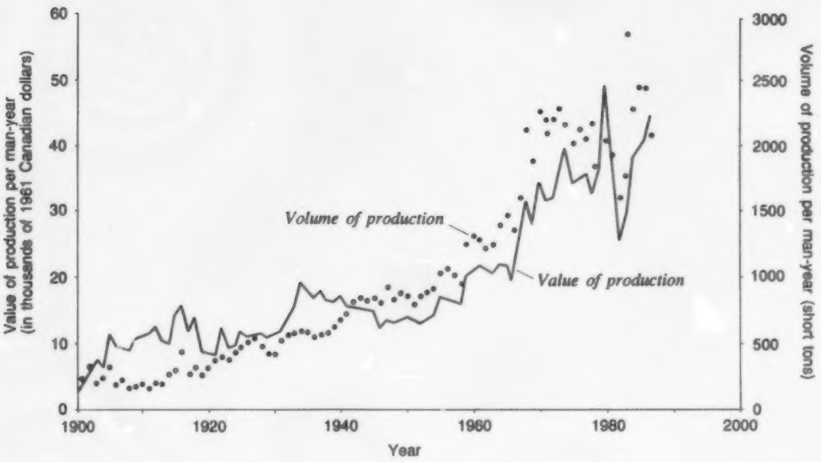


Figure 3.7. Labour productivity in metal mining. Value of production per man-year and volume of production per man-year reveals similar increasing and decreasing trends over the past century.

mineral exploration and development; and 3) a tightening of the legal framework within which the sector must operate, which results in increased "mandated costs". These challenges are not unique to Ontario, they are faced by the mineral sectors of other jurisdictions. The first of these 3 factors is impersonal, market driven, and is not subject to policy influence. The latter 2 "institutional" factors are not expressions of anti-mining sentiment, but result from the necessity to reconcile the interests of society as a whole with those of the mining sector. They are compromises which must be given time for adjustment. These 3 challenges make the mineral sectors' competition for scarce investment capital more complex.

The increasing institutional constraints confronting Ontario's mineral sector typically arise out of the competition of claims and interest in the political arena. They fall into the following major categories: 1) parks policy, 2) native land claims, 3) environmental legislation and taxation, and 4) increases in indirect labour costs resulting from changes in labour legislation. The first 2 of these categories, pose the challenge of a declining land base and result in a reduction in supply of exploration targets. The effect of the other 3 can be summarized as resulting in increased input factor costs, that is, of the factors of production environment, government and labour. Of these challenges, the flat to declining real secular price trend is by far the most important. Challenges posed by declining land base and increasing institutional constraints are rooted in human law and can be eased by policy, whether through modification of severity, timing, or mode of implementation. The price trend challenge, rooted in economic law, is not subject to policy action, but can only be overcome by technological progress.

Declining Price Trend

A composite index of constant dollar metal prices commencing in 1840 (Figure 3.8) shows extraordinary stability both for the long run and for 3 major subperiods (1840–1947, 1947–1974, 1974–1989). We exclude here the 1988 abnormality which is now in the process of correcting itself. The stability of real metal price trends is reflected in the correlation coefficients. Crucial "breaking points" in the trend of real metals prices since 1840, were the years 1947 and 1974, bracketing the abnormal period of consistently rising real prices. General insights can be reached by comparing the metals price index with the wholesale price index.

During the first of the 3 periods—from 1840 to 1947—the metals price index (MPI) trends fairly smoothly downwards, while the wholesale price index (WPI) moves relatively slowly upwards. During these 107 years, a major move from the trend in the wholesale price index is generally associated with one in the opposite direction in the metals price index.

For the period 1947 to 1974, the 2 indices interact differently. The comparatively steep, smooth increase of the WPI from 1947 to 1974 coincides with a significant rise in the MPI. During this period, popular fallacies about impending resource scarcities, which projected ever-increasing mineral resource prices in real terms, and about various future crises appeared and affected decisions.

Beginning in 1971, and particularly after 1974, another change in the relation of the 2 indices becomes obvious. The MPI compensated for the previous aberration and resumed its historical downward drift. The author concludes that the long-term real metal price trend will continue flat to slightly downward sloping into the future in spite of increasing demand and decreasing grades. This conclusion is supported on a theoretical level by the implications of Lancastrian demand theory: consumer preferences do not relate to final products such as specific metals, but to bundles of very general characteristics embodied in those metals, with each metal possessing a number of them. As the consumer cannot purchase characteristics directly, he has to choose between different approximations of his ideal. Materials science keeps forever broadening the menu of materials available, introducing new alloys, plastics and composites. New, improved materials usually enter production processes, not on the basis of price, but on the basis of superior characteristics. When mass consumption is achieved, unit prices decline, similar to the price history that has characterized the introduction of new metals.

Declining Land Base

An important factor in the decrease of Ontario's land base which is open for exploration, is the increase in the area designated as parks, where prospecting and mining are prohibited. Park policy affects a greater area than native land claims and perhaps closes areas to mineral development more effectively. Ontario's recent park policy increased parklands by about 15% to 63 000 km^2. This area covers a square, 251 km on a side, and if superimposed upon southern Ontario, would cover the area bounded by Trenton–Paris–Tobermory–Mattawa. It is 11 times the size of Prince Edward Island. Compared internationally, such an area would be equal in size to West Virginia or one and one half times the size of Switzerland.

Aboriginal land claims have the effect of imposing a "caution" on disputed areas which prohibits further claim staking. If the native land claim is rejected by the courts, then the "caution" is lifted. If the land claim is settled in favour of the aboriginal group, the prospector or developer must negotiate with the aboriginal group in addition to federal and provincial agencies. Based on the evidence of native land claims settled in Canada and in the United States, aboriginal groups are unlikely to be categorically opposed to mineral development. In that case, the area under question may be opened again to mineral activity.

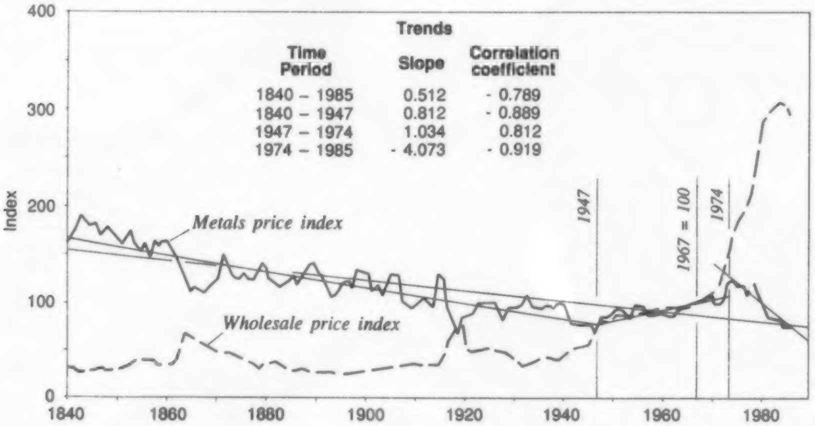
The effects of changes in environmental legislation and taxation upon the future health of the mining industry operate in similar ways (Anders et al. 1980). Increasing or tightening these 2 constraints reduces after-tax profitability of an investment and thus the firm's rate of investment during a particular period of time. The industry's marginal return on investment is then shifted downward and some investments that would have been profitable otherwise will not be made. The relevant determinant here is not the ratio of mandated environmental or tax expenditures to cost, but the ratio of the discounted stream of expenditures to the discounted stream of net returns—a much larger figure. We have estimated the magnitude of such effects (Anders et al. 1980). Such models consider changes in corporate income tax rates, depletion rules, provincial mining taxes, the proportion of corporate income attributable to mining, environmental controls and discount factors. Quantitative estimates of the magnitude of the major changes initiated in Ontario during the 1970s range from 25 to 43% reductions in investment as compared to the levels that would have prevailed had such changes not been implemented.

Over recent decades, the most significant effect for Ontario metal mining has been an increase in the ratio of indirect to direct labour costs from about 5 to 7% immediately after WWII, to about 40% during the mid- to late-1980s. This represents an 8- to 10-fold increase in the gap between the price of labour, the gross wages received by the mineral sector employee, and the cost of that same labour to the employer.

Ontario's mining industry has coped successfully with these significant cost increases. It has overcome them primarily by increasing productivity through substitution of capital for labour, that is by implementing technological change, and by product diversification through research and development.

Increasing Productivity and Product Diversification Through Research and Development

A series of studies conducted under the leadership of Alan Green of Queen's University allow us to highlight, in economic terms, the direction and the magnitude of the changes which occurred (Green and Green 1985, 1987, in press).



Time Period	Slope	Correlation coefficient
1840 – 1985	0.512	- 0.789
1840 – 1947	0.812	- 0.889
1947 – 1974	1.034	0.812
1974 – 1985	- 4.073	- 0.919

Figure 3.8. Metals and wholesale price indices.

In the post-WWII decades, labour productivity as well as total-factor productivity[2] made significant progress. In the early 70s, total-factor productivity began to decline. This was caused primarily by slowdowns in worldwide mineral demand which were not anticipated. The intensity of this decline was among the worst among Canadian industries. Industry did not readily adjust to changing international market conditions because a disproportionate rise in the number of nonproduction workers during the preceding years added to the supply of quasi-fixed factors of production. Furthermore, indirect unit labour costs grew much faster than direct wages. These disproportionate increases were much greater in larger firms and were concentrated in the early- to mid-1970s. In the preceding decade, industry had invested heavily in new capital equipment, increasing its fixed costs of production, and problems arose when the anticipated demand increases and accompanying strong commodity prices failed to materialize through most of the 1970s.

Beginning in the late 70s, the situation started to turn around again, even though productivity growth in the mineral industry continued to be slower than for the economy as a whole. This trend was not due to underinvestment, since annual growth of mineral industry capital stock after 1980 was about equal to that of the country as a whole. The industry was also absorbing the latest technology. The reason for slow growth in total factor productivity was the continuing slow growth in markets, a combination of declining real output and expanding capital stock forced down the growth rate of total-factor productivity. In the labour force, a relatively high growth of white collar workers occurred during the late 60s and into the 70s. This was reversed in the 80s along with a downsizing of the blue collar work force. This led to significant increases in the capital to labour ratio. Although capital costs apparently increased as a result of higher interest charges and steep-rising prices for capital goods, producers were not deterred. The industry was undergoing an investment boom, in anticipation of a resource boom expected in the 90s. While costs of indirect payments per employee continued to rise, the relative rise in costs of indirect wages to direct wages came to a halt. The "pay-off" in improved efficiency to be expected from these recent developments will likely materialize as stronger markets and higher current dollar commodity prices emerge in the 90s.

The level of domestic processing of ores hoisted in Ontario has been increasing dramatically throughout this century and has accelerated significantly during the last 2 decades. Periodic sharp increases in the ratio of unit prices to world metal prices throughout this century indicate that the composition of total production has shifted toward products with higher levels of added value. These shifts were associated both with technological change and with market forces. An indication of the extent to which smelting and refining growth has been due to economic forces following the maturation of Ontario's economy, has been the increase of industries processing raw materials originating outside of Ontario, such as increasingly significant parts of Ontario's steel industry.

Between the 1970s and 1980s, there was an important shift in corporate planning in Ontario's nonferrous metal industry from a production-driven to a market-driven focus. Thus, industry is beginning to transform its production processes to more clearly meet the needs of customers. Over the last 2 decades, a dramatic picture of the increase in value-added and product diversification, (i.e., change in the product-mix) is provided by Ontario's nickel industry. All Ontario refined nickel products were divided into 3 groups of equal volume in ascending order of value-added. Figure 3.9 shows the percentage difference in value per pound between the highest and lowest valued third of production. The growth of this percentage difference from less than 5% in

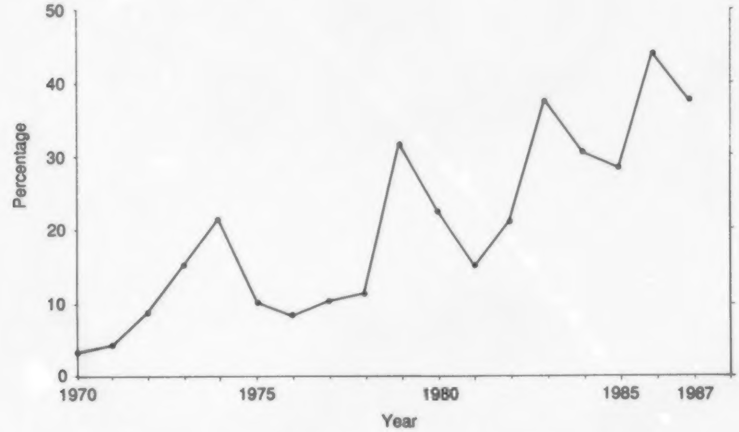


Figure 3.9. Growth of domestic upgrading. The percentage difference between the average unit prices realized for the lowest and highest valued third of Ontario nickel production.

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[2] *Total factor productivity is the difference between the growth of output and the weighted growth of all the inputs where the weights are the factor shares of the inputs. As a residual, total factor productivity captures everything, other than the measured inputs, that influences output and is generally viewed as an index of technological change.*

1970 to about 40% in the late 80s shows the extent to which refined production is increasingly focussed on maximizing value-added.

In addition to the expectations resting in the technology and capital assets base which has been created over the last decade, Ontario's mineral sector in the 1990s will have to continue its commitment to accelerated research and development in prospecting, mining and processing technology and production diversification.

CONCLUSIONS

The central issues which the Ontario mining industry has to face for the future are rooted in the following facts:

1. A healthy Ontario mineral industry must be based on export of metals, that is sales in offshore markets in which we do not affect a world price.
2. World prices of our most important metallic commodities have been declining and the secular trend will continue to do so in real terms.
3. Competition with our metals will increase from offshore sources of the same metals, including sources where government ownership allows major subsidization or control of costs. In the wider materials market, our metals must also compete with ceramics, plastics and composite material substitutes.
4. Most unit factor costs have been increasing and may be expected to continue to do so.

Only a change in expected after-tax rate of return on mine investment will lead to a change in levels of projected investments, and in turn to a change in levels of exploration activity. Among the many factors that affect these expectations at the beginning of this decision chain, Ontario's geology, the location of orebodies, and the world market prices cannot be influenced by corporate or governmental policy. Factors such as the level of geological knowledge, technological and managerial sophistication and entrepreneurship, can be influenced by private and public sector decision-makers. Corporate and government policy must continue to influence the industry in a positive fashion to maintain and enhance Ontario's role on the world's mineral markets and also the role of the mineral sector in the province's economy.

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Superior Province

Chapter 4

Archean Geology of Ontario: Introduction

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The Superior Province is the largest Archean craton in the world with an area of 1 572 000 km² composing 23% of the Earth's exposed Archean crust. The Ontario part covers 565 000 km². The Superior Province is isolated from neighbouring Archean blocks by Proterozoic orogens (Figure 4.1) and the (0.5 to 0.25 Ga) Appalachian Orogen.

The Superior Province differs substantially from many Archean cratons in several respects:

1. **Subprovincial structure:** The Superior Province is characterized by linear, fault-bounded subprovinces, each with distinctive rock types, structures, ages and metamorphic conditions.
2. **Sedimentary subprovinces:** This subprovince type has been considered unique to the Superior Province (e.g., Card 1990). Recent work has shown, however, that sedimentary subprovinces may exist in the Aldan–Stanovik shield (Dook 1989) and the Yilgarn Block of the Australian shield (D. Groves, University of Western Australia, Perth, personal communication, 1990).
3. **Timiskaming-type sequences:** The granite-greenstone subprovinces of the Superior Province include "Timiskaming type" (cf., Thurston and Chivers 1990) greenstones which consist of alluvial-fluvial sedimentary rocks and alkalic to calc-alkalic volcanic rocks unconformably overlying older greenstones. These sequences were deposited after early deformation and prior to late deformation of greenstone belts. Timiskaming-type sequences have been compared to molasse, modern sequences deposited in pull-apart basins (Thurston and Chivers 1990), forearc basins or senescent arcs (Corfu et al. 1991).
4. **Diversity of supracrustal environments:** Recent publications have identified a broad range of depositional settings for Superior Province greenstone sequences. These include: platformal or foredeep, mafic plain or oceanic volcanic, magmatic arc, pull-apart basin (Thurston and Chivers 1990) and back arc basin (*see* Jackson and Fyon, this volume).
5. **Diversity of granitoid types:** In this compilation (*see* Maps 2541 to 2545, map case), 5 major types of granitoid units are recognized, based upon types of petrogenetic suites and tectonic criteria: a) gneissic tonalite suite, b) foliated tonalite suite, c) diorite-monzonite-granodiorite suites, d) muscovite-bearing granites and e) massive granodiorite-granite suites. The Superior Province within Ontario is the largest area over which a comprehensive interpretation of granitoid rocks has been proposed.
6. **Geochronological data base:** Large parts of the Superior Province have been dated by high precision U-Pb zircon geochronology; this provides an unrivalled basis for the interpretation of relationships between the major geological units.

The Superior Province is subdivided into subprovinces–defined below–characterized by distinctive rock types. With a few exceptions, the organization of the Superior Province section of this volume is based on the subdivisions of Card and Ciesielski (1986) in which the various types of subprovinces (Figure 4.2) are described as follows.

1. "Volcano-plutonic" subprovinces consist of low metamorphic grade "greenstone belts" of metavolcanic rocks of various ages surrounded and cut by granitic rocks. The map pattern of greenstone belts was seen as a product of multiple deformational events that produced "sinuous, bifurcating, commonly synformal metavolcanic belts" interrupted by domical gneissic regions.
2. "Metasedimentary" subprovinces were acknowledged to be accumulations of largely clastic metasedimentary rocks with low-grade margins and amphibolite to granulite facies interiors. The origin of these metasedimentary rocks and their relationship to adjoining volcano-plutonic subprovinces was not well defined by Card and Ciesielski (1986).
3. "Gneissic" or "plutonic" subprovinces are characterized by tonalitic gneiss, with extrusive and intrusive mafic enclaves at a variety of scales, and more massive early sodic and later potassic granitoid plutons. Supracrustal rocks are scarce in these subprovinces (cf., Beakhouse 1985).
4. "High-grade gneiss" subprovinces consist of amphibolite to granulite facies meta-igneous and supracrustal gneisses cut by tonalitic, granodioritic and syenitic intrusions (cf., Percival and Card 1985). Typically, the gneissic regions exhibit domal structural patterns with relatively moderate dipping foliations and zones of "straight gneisses" characterized by linear foliation trends and steep dips.

The subdivisions of Card and Ciesielski (1986) have been modified in this volume. "Volcano-plutonic" subprovinces are hereafter referred to as "granite-greenstone" subprovinces. The term "greenstone belt" is defined below. When referring to individual greenstone belts, names have, where possible, been taken from the literature. Thurston (1990) has

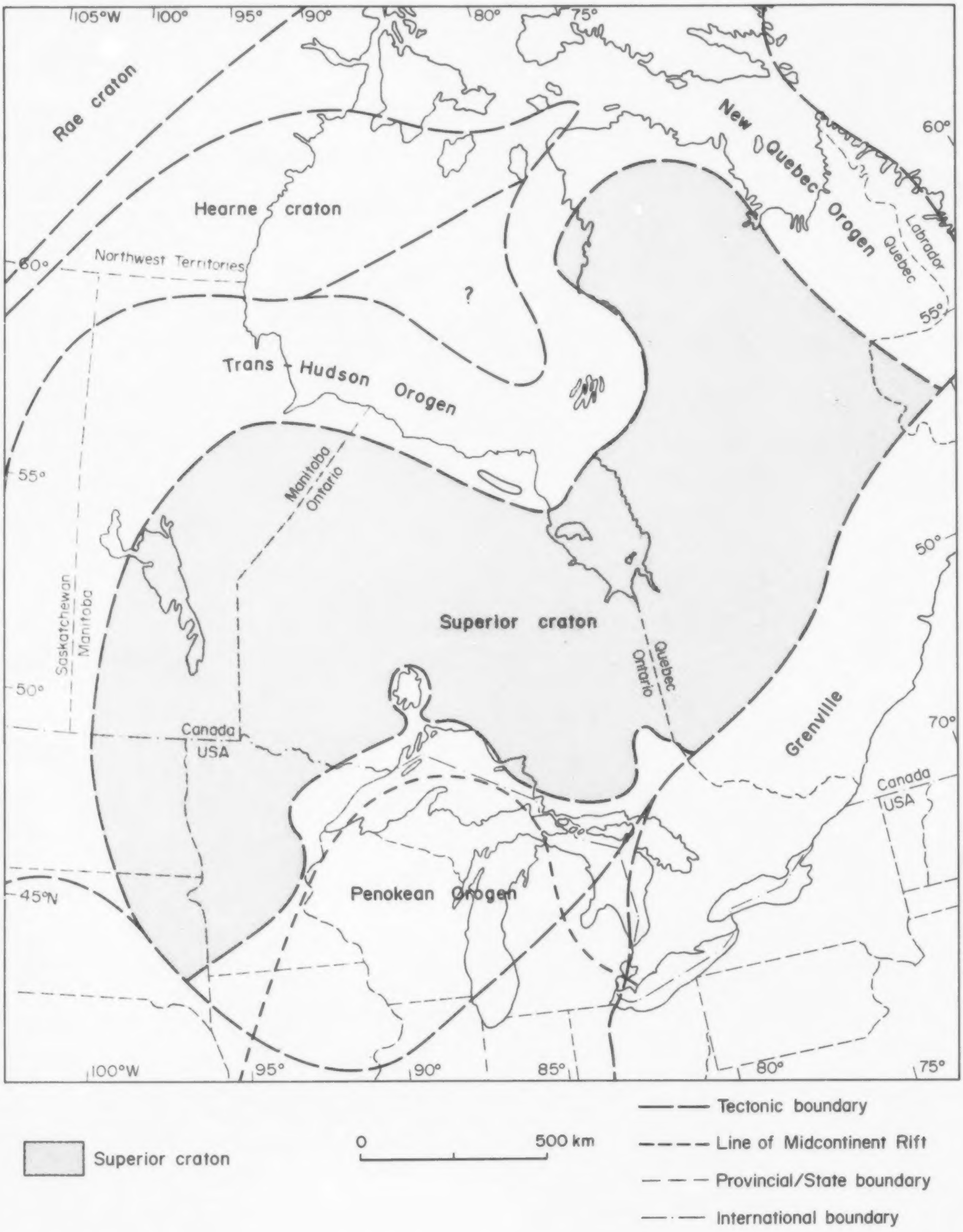
105°W
100°
95°
90°
80°
75°
Rae craton
Hearne craton
Northwest Territories
60°
?
Trans Hudson Orogen
New Quebec Orogen
Labrador
Quebec
55°
60°
Manitoba
Ontario
55°
Saskatchewan
Manitoba
Superior craton
Quebec
Ontario
50°
50°
Grenville
70°
Canada
USA
Canada
USA
Penokean Orogen
45°N
100°W
95°
90°
85°
80°
75°
Superior craton
0
500 km
Tectonic boundary
Line of Midcontinent Rift
Provincial/State boundary
International boundary

Figure 4.1. The Superior Province relative to other North American Archean cratons and surrounding Proterozoic orogens (*modified from* Hoffman 1989).

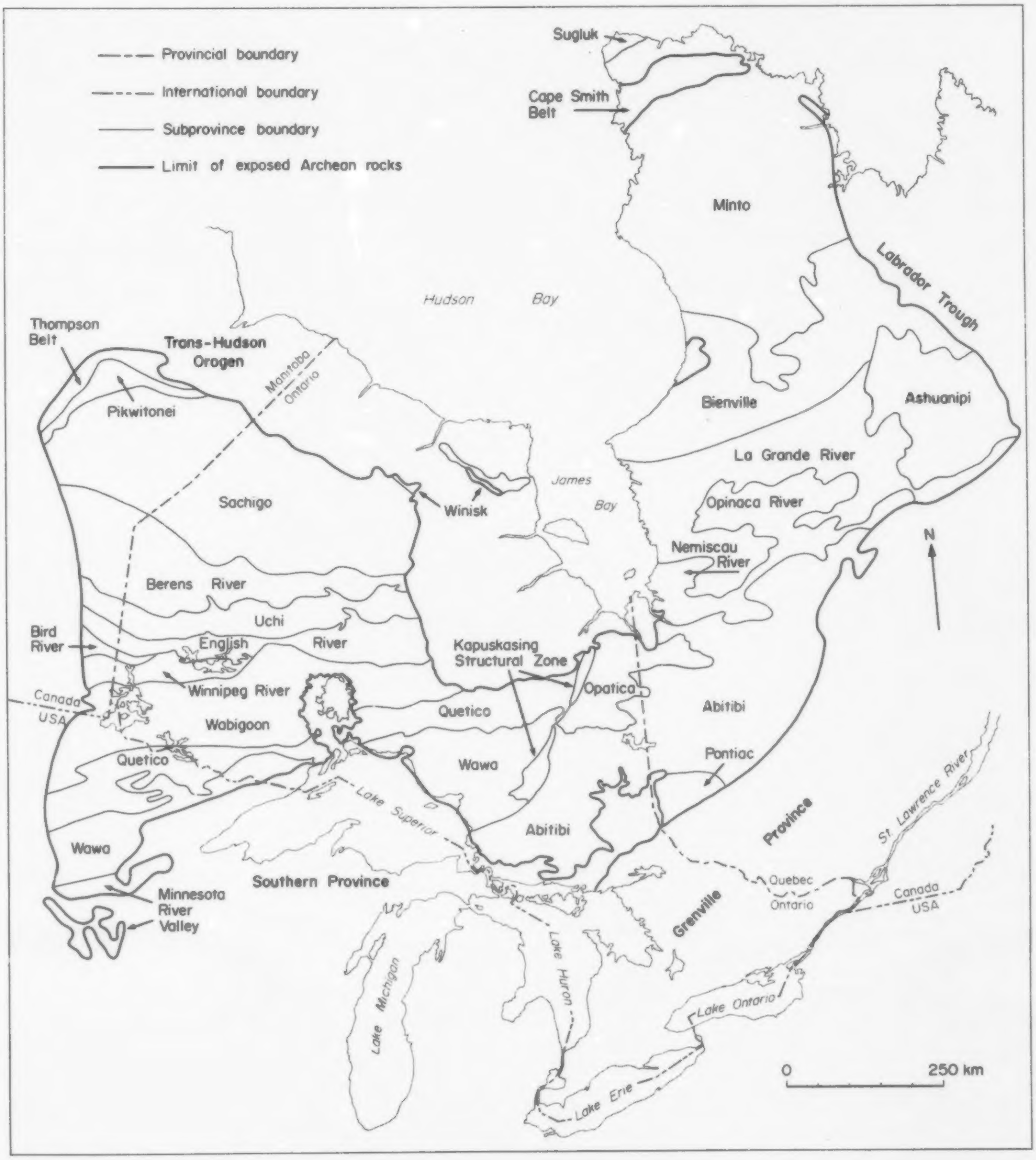
Provincial boundary
International boundary
Subprovince boundary
Limit of exposed Archean rocks
Sugluk
Cape Smith Belt
Minto
Labrador Trough
Hudson Bay
Thompson Belt
Trans-Hudson Orogen
Manitoba
Ontario
Pikwitonei
Bienville
Ashuanipi
La Grande River
Sachigo
Winisk
James Bay
Opinaca River
Nemiscau River
N
Berens River
Uchi
Bird River
English River
Kapuskasing Structural Zone
Winnipeg River
Opatica
Canada
USA
Wabigoon
Quetico
Abitibi
Quetico
Wawa
Pontiac
Lake Superior
St. Lawrence River
Province
Abitibi
Wawa
Minnesota River Valley
Southern Province
Grenville
Quebec
Ontario
Canada
USA
Lake Huron
Lake Michigan
Lake Ontario
0
250 km
Lake Erie

Figure 4.2. Subdivision of the Superior Province into subprovinces (*modified from* Card and Ciesielski 1986).

suggested that high-grade gneissic or granulite subprovinces are high-ranking metamorphic equivalents to the subprovince types previously listed. Therefore, "high-grade gneiss subprovinces" such as the Kapuskasing Structural Zone (Card and Ciesielski 1986) are not described as separate subprovinces. Rather, the various parts of this zone of high-ranking metamorphic rocks are described in this volume as extensions of the Wawa granite-greenstone and Quetico metasedimentary subprovinces (*see* Williams, Stott, Thurston et al., this volume).

The subprovincial classification of Card and Ciesielski (1986) has been widely accepted. Therefore, the Superior Province section of this volume is organized into chapters which describe each of the subprovinces as they were defined by Card and Ciesielski (1986). Each subprovince was assigned to an author or team of authors.

Variations in the types of information available for each subprovince, in concert with contrasts in geological experience of the various authors, have produced inevitable differences between the individual chapters that form the Superior Province section.

1. The chapter on the northwestern Superior Province (*see* Thurston, Osmani et al., this volume) incorporates elements of the Uchi, Berens River and Sachigo subprovinces of Card and Ciesielski (1986). The chapter emphasizes tectonic analysis of the area, based on new structural data and several new U-Pb zircon ages (D.W. Davis, Royal Ontario Museum, personal communications, 1989, 1991). These data have permitted the establishment of new tectonic subdivisions.

2. The Uchi Subprovince chapter (*see* Stott and Corfu, this volume) emphasizes the large-scale structural and geochronologic evidence for an accretionary origin of the supracrustal sequences. Interpretations of the tectono-stratigraphy are supported by aeromagnetic and electromagnetic data. The chapter also develops models to explain the effect of various episodes of intrusive activity on the development of structures in the supracrustal rocks.

3. The English River Subprovince chapter (*see* Breaks, this volume) discusses the evolution of a sedimentary subprovince with a petrologic and metamorphic emphasis, incorporating new geochronological data (F. Corfu, Royal Ontario Museum, personal communication, 1990).

4. The Winnipeg River Subprovince chapter (*see* Beakhouse, this volume) is a petrologically comprehensive description of a plutonic subprovince. Geochronology combined with thorough descriptions of units are used to substantiate the author's interpretations.

5. The Wabigoon Subprovince chapter (*see* Blackburn et al., this volume) provides a comprehensive compilation of all data presently available on the supracrustal rocks of the subprovince. A new structural synthesis has been supported by several new geochronological determinations by D.W. Davis, Royal Ontario Museum (*see* Blackburn et al., this volume).

6. The Quetico Subprovince description (*see* Williams, this volume) documents the structural development of a sedimentary subprovince and its relationship in structural and stratigraphic terms to the neighbouring Wabigoon and Wawa subprovinces. Thus, it provides a more complete documentation of the evidence for an accretionary prism origin of the subprovince than is suggested in earlier work (Percival and Williams 1989).

7. In the Western Abitibi Subprovince chapter (*see* Jackson and Fyon, this volume), several new depositional environments are identified and new geochronological data are used to support a thorough re-evaluation of the most intensely explored granite-greenstone subprovince in Canada. The structural synthesis provides new insight into assemblage relationships and demonstrates the complexities within what was once considered just one very large greenstone belt.

8. The Wawa Subprovince chapter (*see* Williams, Stott, Heather et al., this volume) postulates tectono-stratigraphic relationships between the isolated greenstone belts throughout the subprovince. The stratigraphy of the western part of the subprovince is correlated with that in the Minnesota part of the subprovince.

New structural and geochronological data incorporated into this volume and the accompanying maps have brought about substantial revisions to Card and Ciesielski's (1986) Superior Province subdivisions of northwestern Ontario. The revised tectonic subdivisions are based upon the recent recognition of major shear zones cutting the Sachigo Subprovince (cf., Osmani and Stott 1988; *see also* Map 2584, map case), remapping of greenstone belts in northwestern Ontario (Thurston et al. 1987; Cortis et al. 1988), and numerous new geochronological determinations in northwestern Ontario by D.W. Davis (*see* Maps 2575 and 2576, map case) and in the Abitibi Subprovince by F. Corfu (*see* Jackson and Fyon, this volume). The basis for new tectonic subdivisions of the Superior Province in northwestern Ontario is described in the chapters on the northwestern Superior Province (*see* Thurston, Osmani et al., this volume) and the Uchi Subprovince (*see* Stott and Corfu, this volume).

The substantial changes in our understanding of the Superior Province developed during this compilation, and the length of the Superior Province section have necessitated a new synthesis. This is given in a tectonic summary (*see* Williams, Stott, Thurston et al., this volume) where the newly defined tectonic subdivisions of the Superior Province are presented.

The chapters on individual subprovinces and the tectonic summary of the Superior Province synthesize data from a large part of a vast Archean craton. The approach in each chapter has involved description using nongenetic terminology followed by a discussion of models. We hope that the descriptive part of the volume will stand the test of time. Discussions of the more speculative, model-based interpretations may serve to stimulate development of comparable models for other Archean cratons.

NOMENCLATURE

Our understanding of the Superior Province has been advanced by the application of descriptive terminology used in describing modern orogens. The terminology used in the Superior Province chapters is reviewed in the next paragraphs.

Subprovinces are composed of a single terrane, for example, the Quetico Subprovince (*see* Williams, this volume), or several terranes, such as the Abitibi (*see* Jackson and Fyon, this volume) or Wabigoon (*see* Blackburn et al., this volume) subprovinces. There is emerging evidence (*see* Thurston, Osmani et al., this volume; Stott and Corfu, this volume; Williams, Stott, Thurston et al., this volume) for multiple orogenic events in the Superior Province. The major orogenic event in the Superior Province (cf., Okulitch 1988) has been known as the "Kenoran orogeny". To avoid potential confusion, this term and others are defined in the following section.

DEFINITIONS

Terrane: a fault-bounded package of strata that is allochthonous to, and has a geological history distinct from, the adjoining geologic units (e.g., Howell 1989, p.214).

Superterrane: a composite terrane, consisting of 2 or more component terranes that were amalgamated prior to their accretion to other terranes and superterranes during the Kenoran orogeny. Superterranes generally coincide with the subprovinces. The Wawa superterrane is an example having formed as an amalgamation of terranes, at least in large part, prior to accretion of other granite-greenstone superterranes and sedimentary terranes to form the Superior Province.

Tectonic assemblage: as informally applied in the Ontario part of the Superior Province, an assemblage consists of stratified volcanic and/or sedimentary rock units built during a discrete interval of time in a common depositional or volcanic setting. The rock units typically share a common or dominant lithofacies; they may also share some additional attributes, such as structural, metamorphic, geochemical and geophysical features. The units within an assemblage may display features of primary stratification, yet they need not conform to the Law of Superposition. An assemblage is typically bounded by faults, unconformities or intrusions. An assemblage may consist of one or more groups or formations and one or more assemblages may constitute a terrane. Some Archean assemblages are closely comparable to other assemblages of different ages in different parts of the Superior Province and may represent a common tectonic setting that recurs in time and space across the Superior Province. However, at present in Ontario, our assemblage subdivisions of the Precambrian crust anticipate, but do not explicitly define, tectonic settings comparable to modern analogues. Revisions of assemblages, as herein defined, should progressively incorporate interpretation in terms of modern tectonic analogues.

(Fabric) domain: a volume of rock, bounded by compositional or structural discontinuities, within which there is structural homogeneity (cf., Bates and Jackson 1987). This term can apply to an entire subprovince (e.g., Sachigo Subprovince?) or just an assemblage.

Greenstone belt: elongate or belt-like, kilometre-scale areas of supracrustal rocks within granite-greenstone subprovinces, with tectonic or intrusive boundaries. Greenstone belts may consist of one or more assemblages. A subprovince generally consists of several greenstone belts.

Subprovince: a fault-bounded, medium- to large-scale region characterized by similar rock types, structural style, isotopic age, metamorphic grade and mineral deposits.

Kenoran Orogeny: an event of folding and faulting, including thrusting, accompanied by metamorphism and plutonism, that constitutes the culminating event in the progressive accretionary history forming the Superior Province craton. Tectonism was most intense from about 2720 to 2660 Ma.

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Chapter 5

Northwestern Superior Province: Review and Terrane Analysis

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Abstract

This chapter describes the northwestern Superior Province, including the Sachigo, Winisk and Berens River subprovinces and the northwestern part of the Uchi Subprovince of previously proposed tectonic subdivisions. The region is composed of linear to arcuate greenstone belts surrounded by felsic plutonic rocks and gneisses. Greenstone belts strike in a southeasterly direction in the north, and in a northerly direction in the southwest part of the region. Greenstone belts in the central part of the northern Superior Province are composed of rare 3.02 to 2.98 billion-year-old mafic to felsic (arc) greenstones overlain by widely distributed quartz-rich metasedimentary rocks containing 2.9 to 3.0 billion-year-old detrital zircons and intercalated with subordinate metavolcanic rocks. The quartz-rich metasedimentary sequences occur in all major greenstone belts in the central part of the northwestern Superior Province. The extent of the quartz-rich sequences, and their provenance and ages, suggest that erosion of extensive 2.9 to 3.0 billion-year-old sialic crust was involved in their genesis. Primary structures and stromatolite-bearing units indicate that these sequences represent examples of sediments deposited on shallow-water epicratonic platforms, in submarine fans, and possibly in wrench basins. The quartz-rich sequences are overlain by younger resedimented turbidite associations and mafic metavolcanic units (oceanic volcanism?).

Greenstone belts in the northern part of the region are composed of 2.8 and 2.7 billion-year-old mafic to felsic metavolcanic (arc) rocks with a lesser component of alluvial-fluvial metasedimentary rocks and alkalic metavolcanic rocks spatially associated with major shear zones. The latter units are geochemically similar to modern shoshonitic sequences which develop during the late stages of arc magmatism.

Felsic plutonic rocks in the region include, in order of age, based on field relations: a gneissic tonalite suite, a foliated tonalite suite, massive granodiorite to granite suite, muscovite-bearing granite suite, and a diorite-monzonite-granodiorite suite. The gneissic tonalite suite includes abundant mafic inclusions representing supracrustal or tectonized dike enclaves and, in part, forms pre-greenstone basement. The foliated tonalite suite is commonly early tectonic in age, and the remaining granitoid units are syntectonic to posttectonic in age.

Throughout the northwestern Superior Province, greenstone belts have been subjected first to a recumbent folding and thrusting event, and then to an upright folding and shearing event. This last event included the intrusion of major batholiths, and development of extensive shear zones.

The range in the ages of volcanism (3.02 to 2.72 Ga) and posttectonic plutonism (2.87 to 2.71 Ga), and the variation in character and distribution of supracrustal units in the northwestern Superior Province, suggests that the previously proposed tectonic subdivisions require revision. We define 6 shear-zone–bounded terranes based on age of volcanism, age of posttectonic plutonism and inferred age of activity on the boundary shear zones. The ages of the terranes, and their ages of juxtaposition, become younger north and south from a central core of felsic plutonic rocks and quartz-rich greenstones.

INTRODUCTION

This chapter describes the northwestern part of the Archean Superior Province including the Sachigo, Berens River and Winisk subprovinces described by Card and Ciesielski (1986) (Figure 5.1). The northwestern Superior Province is bordered on the northwest by the Thompson Belt, a zone of granitoid gneisses and lesser ultramafic and mafic intrusions forming the collisional zone between the Superior Province and the 1.8 billion-year-old Trans–Hudson Orogen. The western limit of the northern Superior Province, beneath a thin Paleozoic cover sequence in Manitoba, is the southward extension of the boundary between the Trans–Hudson Orogen and the Superior Province. The eastern boundary is the Winisk River fault (Hoffman 1989, p.469), separating the northwestern Superior Province from the Fox River Belt of the Trans–Hudson Orogen. Superior Province rocks of the Winisk Subprovince are exposed within the Sutton Inlier. The southern boundary has not been clearly delimited and is discussed in a subsequent section. We summarize the major characteristics of the northwestern Superior Province in the following paragraphs, concluding with justification for a subdivision of the northwestern part of the Superior Province into several newly defined terranes similar to the subprovinces of the southern part of the Superior Province.

The Sachigo and Berens River subprovinces described by Card and Ciesielski (1986), the northernmost granite-greenstone subprovinces in Ontario, are made up of relatively isolated greenstone belts surrounded by extensive granitic and gneissic units (Figure 5.2). New geochronological, stratigraphic and metamorphic data suggest the subdivisions listed above require revision, as summarized below:

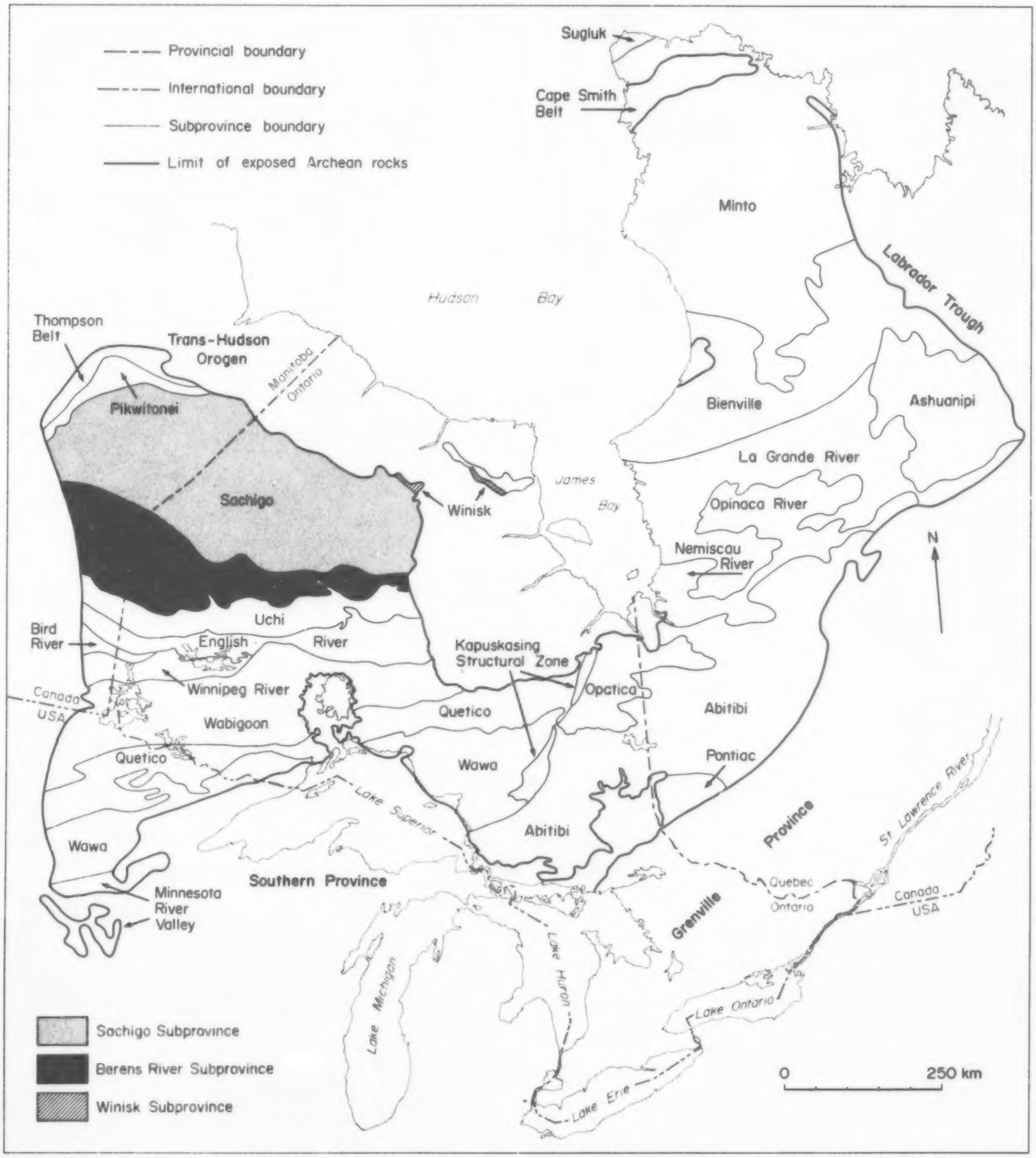

Provincial boundary
International boundary
Subprovince boundary
Limit of exposed Archean rocks
Sugluk
Cape Smith Belt
Minto
Labrador Trough
Hudson Bay
Thompson Belt
Trans-Hudson Orogen
Manitoba
Ontario
Pikwitonei
Bienville
Ashuanipi
La Grande River
James Bay
Sachigo
Winisk
Opinaca River
Nemiscau River
N
Uchi
Bird River
English River
Kapuskasing Structural Zone
Winnipeg River
Opatica
Canada
USA
Quetico
Abitibi
Wabigoon
Quetico
Wawa
Pontiac
Lake Superior
Abitibi
Province
St Lawrence River
Wawa
Southern Province
Minnesota River Valley
Quebec
Ontario
Canada
USA
Grenville
Lake Michigan
Lake Huron
Lake Ontario
Sachigo Subprovince
Berens River Subprovince
Winisk Subprovince
0
250 km
Lake Erie

Figure 5.1. Generalized map of the Superior Province showing location of the Sachigo, Winisk and Berens River subprovinces described by Card and Ciesielski (1986) relative to the remainder of the Superior Province.

1. Greenstone belts in southern Sachigo Subprovince include a) unusual sequences containing quartz-rich conglomerates and sandstones with minor metavolcanic units and b) mafic to felsic volcanic cycles. The quartz-rich sequences are the oldest widespread greenstones in the Superior Province, containing detrital zircons dated at 2.9 to 3.0 Ga, and having a depositional age in the same range (Corfu and Wood 1986; deKemp1987; D.W. Davis, Royal Ontario Museum, personal communication, 1989). They represent a platformal (Wood et al. 1986; Thurston and Chivers 1990) submarine fan or wrench basin setting (Cortis et al. 1989). The granitic provenance of these quartz-rich sedimentary rocks, their extent within the southern part of the Sachigo Subprovince, and the age of detrital zircons within them suggest that the erosion of a widespread, approximately 3.0 billion-year-old sialic area formed these sequences. The ages of greenstones in the Sachigo Subprovince, relative to the younger greenstone belts of the southern Superior Province, suggest that the southern part of the Sachigo Subprovince is, in large part, a remnant of early sialic crust which is not preserved on this scale elsewhere in the Superior Province.
2. Lithologic and geochronological evidence presented below (*see also* Stott and Corfu, this volume) suggests that substantial revisions to the extent of the Berens River Subprovince (as described by Card and Ciesielski 1986) are warranted. Continental arc subprovinces such as the Berens River Subprovince have been considered alternatively as microcontinental cores to arc subprovinces or deeply eroded arc complexes (Hoffman 1989); in the absence of geochronological evidence for the former, evidence for the latter interpretation is presented.
3. The similarities between supracrustal and plutonic units, structural trends, and the ages of units suggest that the southern part of the Sachigo Subprovince, the Berens River Subprovince, and the northwestern part of the Uchi Subprovince (as described by Card and Ciesielski 1986) should be considered a newly defined Uchi–Sachigo–Berens River superterrane.
4. In the northern part of the Sachigo Subprovince (Card and Ciesielski 1986), supracrustal and intrusive rocks display a great range of ages: greenstones range from 3 to 2.7 Ga and posttectonic granitic units range from 2871 Ma (D.W. Davis, Royal Ontario Museum, personal communication, 1990) to 2710 Ma (Corfu and Ayres, in press). Late, unconformable greenstone sequences range in age from 2731 Ma (Corfu and Wood 1986) to 2709 Ma (MEM 1987). Several major shear zones (Osmani and Stott 1988) define blocks containing stratigraphically and geochronologically similar greenstone sequences, forming the basis for the subdivision of the northern part of the Sachigo Subprovince into several newly defined granite-greenstone terranes.

As outlined in the introduction to the Superior Province, assemblages are groups of rock types deposited in specific settings. Assemblages may be in either depositional or tectonic contact with neighbouring assemblages, and together make up a greenstone belt. Descriptions of assemblages form the basis of the following sections, which systematically review the greenstone belts of the northwestern Superior Province. Exposure of these greenstone belts varies from less than 1% to a maximum of 25% in this part of Ontario. Many parts of the area described in this chapter have been mapped only by helicopter-supported reconnaissance; therefore, geological and geochronological detail is not equivalent to that available for areas farther south.

Experience gained by studying the southern part of the Superior Province indicates that it is possible to correlate greenstone sequences over tens to hundreds of kilometres on either side of batholithic complexes, for example, in the Uchi and Wabigoon subprovinces. Northwestern Superior Province assemblages may also be correlated, in some instances over substantial distances, with those at localities for which U–Pb ages are available. Granitoid rocks have been mapped mostly in reconnaissance fashion; the descriptions which follow will be of a general nature.

The descriptive material in the following section is followed by a generalized discussion of structural geology. The concluding section will be a terrane analysis and a tectonic synthesis, proposing a new tectonic subdivision of the northwestern Superior Province.

Mapping History

Mapping in the region progressed from the early reconnaissance investigations (e.g., Bell 1872), to concentration on individual greenstone belts (e.g., Hurst 1930; Satterly 1941) in the 1920s and 1930s. These were followed by helicopter-supported reconnaissance surveys of the entire northern part of the Superior Province (Bennett and Riley 1969; Thurston and Carter 1970; Thurston et al. 1979). Mapping of individual greenstone belts has continued at a detailed scale (e.g., Ayres 1974; Breaks et al., in prep.). Correlation of rocks within greenstone belts and intrusive rocks has been aided by a series of U–Pb zircon geochronological studies carried out in the northwestern Superior Province, which are referred to as an essential part of the lithologic descriptions.

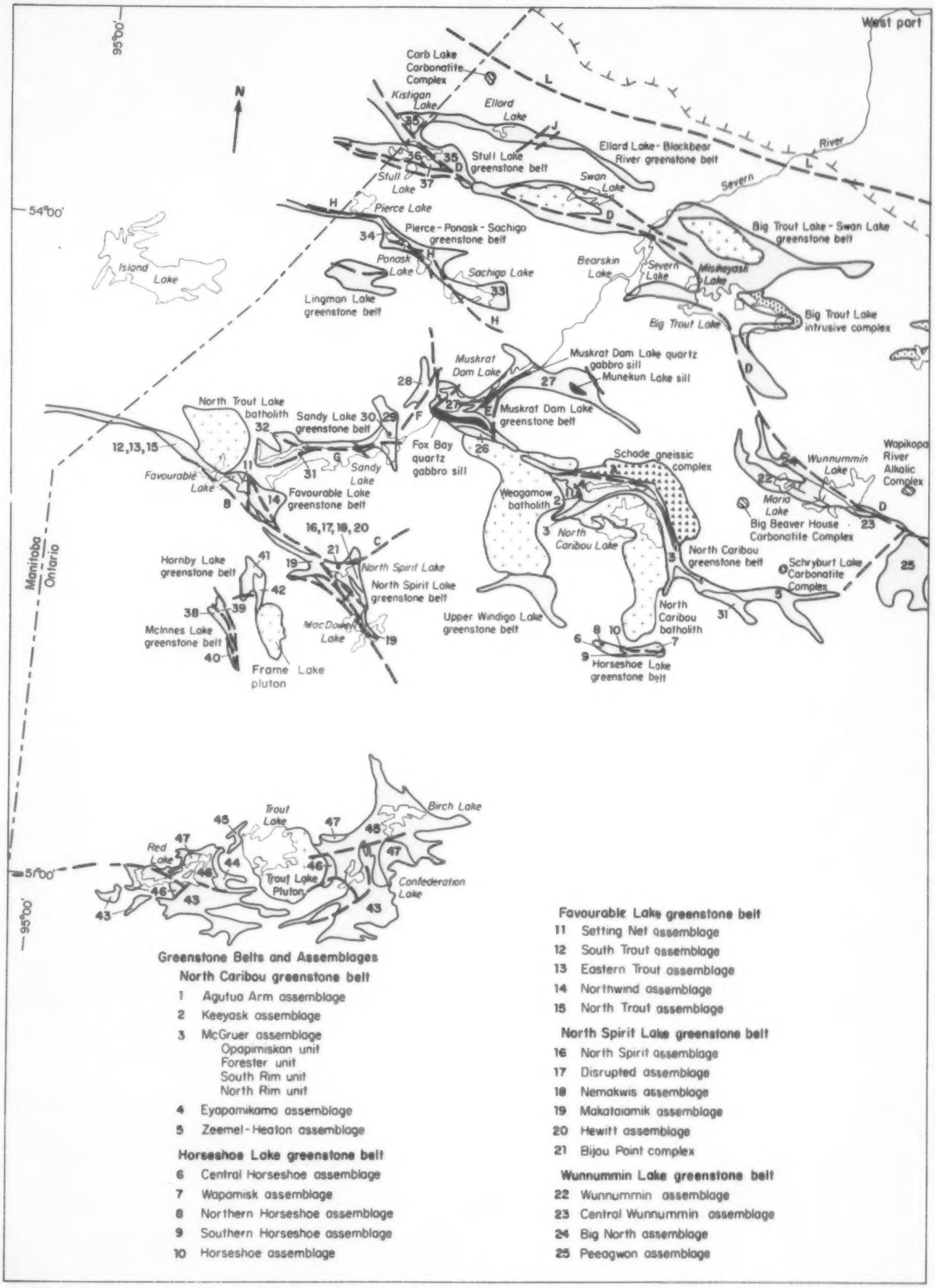
The Bedrock Geology of Ontario maps (*see* Maps 2541 to 2543, map case) portray the smaller greenstone belts and septa of supracrustal rocks between major greenstone belts.

SACHIGO SUBPROVINCE GREENSTONE BELTS

North Caribou Greenstone Belt

The North Caribou greenstone belt is located in the southern part of the Sachigo Subprovince (as described by Card and Ciesielski 1986)(*see* Figure 5.2). The belt consists of 4 supracrustal assemblages (Figure 5.3) modified after Breaks et al. (in prep.) and listed in Table 5.1.

The 2980 million-year-old Agutua Arm assemblage (Davis, in deKemp 1987) represents arc volcanism and is unconformably overlain by quartz-rich metasedimentary


West part
95°00'
54°00'
51°00'
95°00'
N
Manitoba
Ontario
Carb Lake Carbonatite Complex
Kistigan Lake
Ellard Lake
Ellard Lake - Blackbear River greenstone belt
Stull Lake greenstone belt
Stull Lake
Swan Lake
Severn River
Pierce Lake
Pierce - Ponask - Sachigo greenstone belt
Ponask Lake
Sachigo Lake
Island Lake
Lingman Lake greenstone belt
Bearskin Lake
Severn Lake
Mishkeyask Lake
Big Trout Lake - Swan Lake greenstone belt
Big Trout Lake
Big Trout Lake intrusive complex
Muskrat Dam Lake
Muskrat Dam Lake quartz gabbro sill
Munekun Lake sill
Muskrat Dam Lake greenstone belt
Fox Bay quartz gabbro sill
North Trout Lake batholith
Sandy Lake greenstone belt
Sandy Lake
Favourable Lake
Favourable Lake greenstone belt
Schade gneissic complex
Weagamow batholith
North Caribou Lake
North Caribou greenstone belt
North Caribou batholith
Upper Windigo Lake greenstone belt
Horseshoe Lake greenstone belt
Wunnummin Lake
Wapikopa River Alkalic Complex
Maria Lake
Big Beaver House Carbonatite Complex
Schryburt Lake Carbonatite Complex
Hornby Lake greenstone belt
McInnes Lake greenstone belt
Frame Lake pluton
North Spirit Lake
North Spirit Lake greenstone belt
MacDowell Lake
Red Lake
Trout Lake
Trout Lake Pluton
Birch Lake
Confederation Lake
B
C
D
E
F
G
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Greenstone Belts and Assemblages

North Caribou greenstone belt

1 Agutua Arm assemblage
2 Keeyask assemblage
3 McGruer assemblage
  Opapimiskan unit
  Forester unit
  South Rim unit
  North Rim unit
4 Eyapamikama assemblage
5 Zeemel - Heaton assemblage

Horseshoe Lake greenstone belt

6 Central Horseshoe assemblage
7 Wapamisk assemblage
8 Northern Horseshoe assemblage
9 Southern Horseshoe assemblage
10 Horseshoe assemblage

Favourable Lake greenstone belt

11 Setting Net assemblage
12 South Trout assemblage
13 Eastern Trout assemblage
14 Northwind assemblage
15 North Trout assemblage

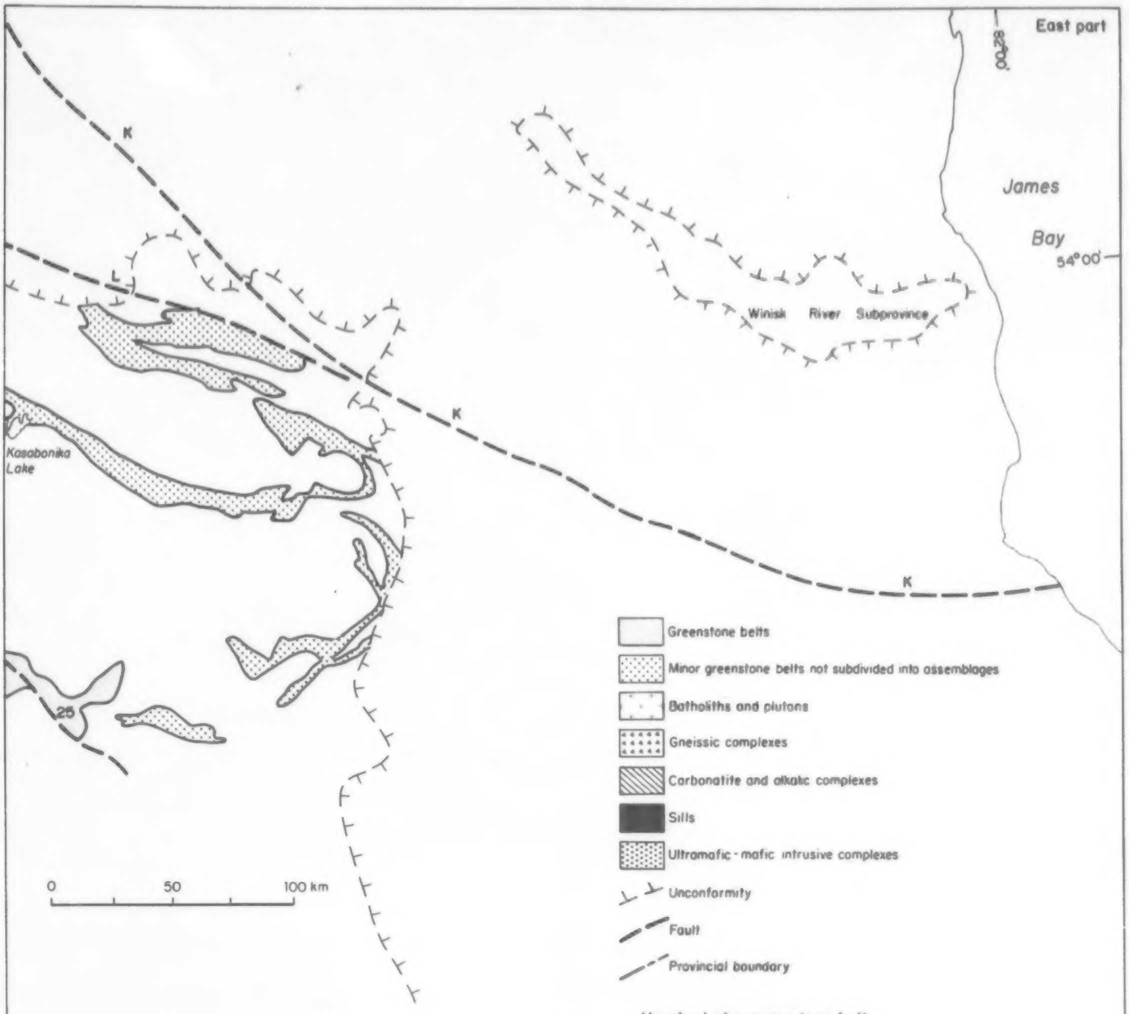
North Spirit Lake greenstone belt

16 North Spirit assemblage
17 Disrupted assemblage
18 Nemakwis assemblage
19 Makataiamik assemblage
20 Hewitt assemblage
21 Bijou Point complex

Wunnummin Lake greenstone belt

22 Wunnummin assemblage
23 Central Wunnummin assemblage
24 Big North assemblage
25 Peeagwon assemblage

Figure 5.2. Generalized map of the Berens River, Winisk and Sachigo subprovinces showing the location of major greenstone belts, batholiths and shear zones based upon the Bedrock Geology of Ontario maps (*see* Maps 2541 to 2543, map case).


East part
87°00'
James
Bay
54°00'
K
K
K
L
Winisk River Subprovince
Kasabonika Lake
25
0
50
100 km
Greenstone belts
Minor greenstone belts not subdivided into assemblages
Batholiths and plutons
Gneissic complexes
Carbonatite and alkalic complexes
Sills
Ultramafic - mafic intrusive complexes
Unconformity
Fault
Provincial boundary
51°00'
82°00'

Muskrat Dam Lake greenstone belt

26	Nekence assemblage
27	Muskrat Dam assemblage
28	Rottenfish assemblage

Sandy Lake greenstone belt

29	North Sandy assemblage
30	Keewaywin assemblage
31	Sandborn assemblage
32	West Arm assemblage

Pierce - Ponask - Sachigo greenstone belt

33	Sachigo assemblage
34	Pierce assemblage

Stull Lake greenstone belt

35	Stull assemblage
36	Rapson assemblage
37	Richardson Arm assemblage

McInnes Lake greenstone belt

38	Western McInnes assemblage
39	Power assemblage
40	Southern McInnes assemblage

Hornby Lake greenstone belt

41	Western Hornby assemblage
42	Findlay assemblage

Red Lake - Confederation Lake greenstone belt

43	Confederation assemblage
44	Bruce Channel assemblage
45	Woman assemblage
46	Balmer assemblage / Ball assemblage
47	Metasediments (2800 - 2900 Ma)

Faults

A	North Rim fault
B	Bear Head fault zone
C	North Spirit Lake shear zone
D	Stull Lake - Wunnummin Lake fault zone
E	Windigo River shear zone
F	Rottenfish Shear zone
G	Sandy Lake shear zone
H	Ponask Lake shear zone
J	South Kenyon Fault Zone
K	Winisk River fault
L	Kenyon Structure

Figure 5.2. Generalized map of the Berens River, Winisk and Sachigo subprovinces showing the location of major greenstone belts, batholiths and shear zones based upon the Bedrock Geology of Ontario maps (*see* Maps 2541 to 2543, map case).

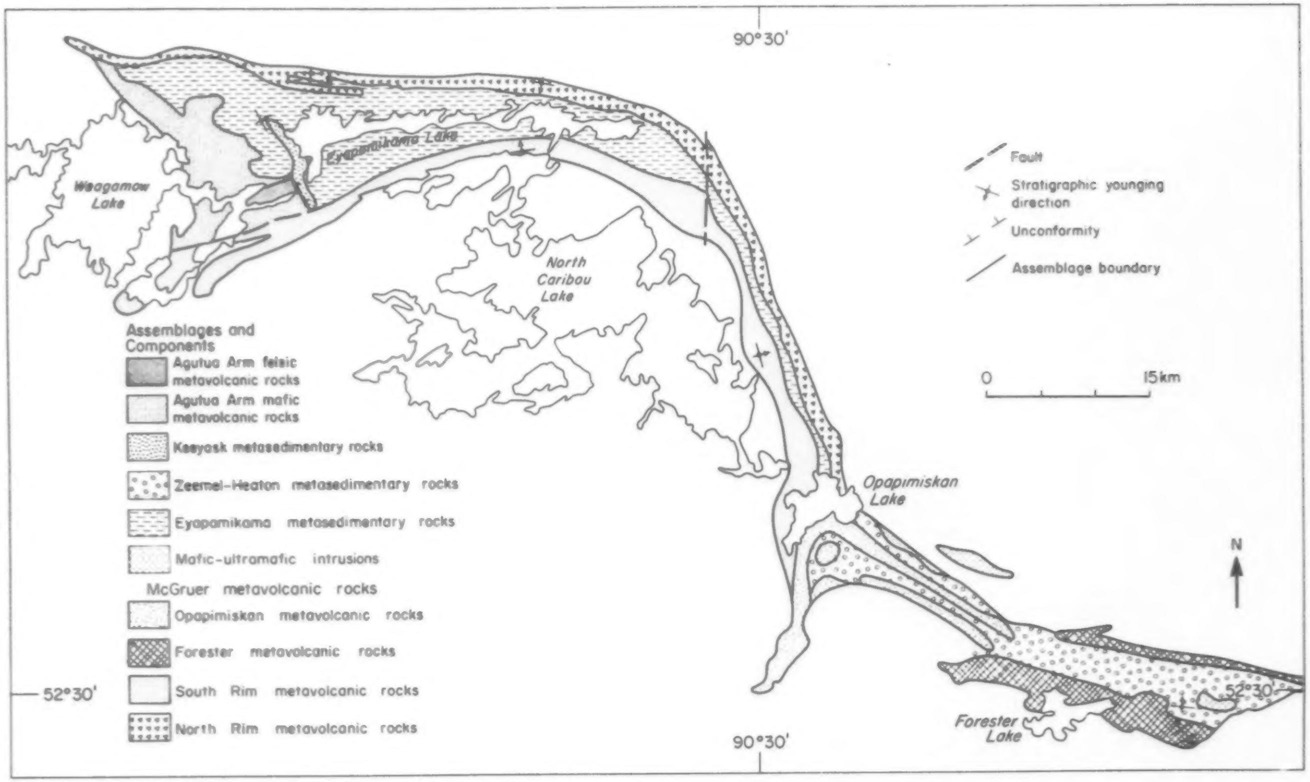

90°30'
90°30'
52°30'
52°30'
Weagamow Lake
Eyapamikama Lake
North Caribou Lake
Opapimiskan Lake
Forester Lake
Fault
Stratigraphic younging direction
Unconformity
Assemblage boundary
0
15km
N
Assemblages and Components
Agutua Arm felsic metavolcanic rocks
Agutua Arm mafic metavolcanic rocks
Keeyask metasedimentary rocks
Zeemel-Heaton metasedimentary rocks
Eyapamikama metasedimentary rocks
Mafic-ultramafic intrusions
McGruer metavolcanic rocks
Opapimiskan metavolcanic rocks
Forester metavolcanic rocks
South Rim metavolcanic rocks
North Rim metavolcanic rocks

Figure 5.3. North Caribou greenstone belt supracrustal assemblages based upon mapping carried out by Breaks et al. (in prep.).

rocks of the Keeyask assemblage which contain 2980 million-year-old detrital zircons. Therefore, based on petrographic and isotopic evidence, the platformal metasedimentary rocks are of magmatic arc provenance (deKemp 1987; Cortis et al. 1989). A shallow-water depositional setting for the basal clastic and carbonate metasedimentary rocks of the Keeyask assemblage is indicated by stromatolites in siltstones and herringbone, planar, trough and hummocky cross-beds in the coarser clastic metasediments (Figure 5.4). The primary structures and the spatial association of quartz arenites and carbonates indicate that this part of the assemblage was probably deposited in a platformal environment (Wood et al. 1986; deKemp 1987; Thurston and Chivers 1990). The upward transition to siltstones and iron formation is consistent with a deepening basin or a lack of sediment supply. This set of environments compares favourably with the foredeep setting described by Hoffman (1987). The upper magnesium-rich volcanic part of the assemblage suggests late rifting during the final stage of assemblage development. The Keeyask assemblage is correlated with the Lundmark Lake metavolcanic formation and the Zeemel–Heaton assemblage on lithologic and structural grounds (Breaks et al., in prep.). The McGruer assemblage comprises the North Rim metavolcanic (2932 Ma) (D.W. Davis, Royal Ontario Museum, personal communication, 1991), the South Rim metavolcanic, the Forester metavolcanic and the Opapimiskan metavolcanic units, based on lithologic and structural correlations suggested by Breaks et al. (in prep.).

At the west end of the belt, the North Rim and South Rim components of the McGruer assemblage are in both depositional and tectonic contact with the underlying Keeyask assemblage (*see* Table 5.1). The North Rim unit is in tectonic and/or intrusive contact with the Schade gneissic complex to the east. The North and South Rim components are correlated on lithologic and structural evidence. These assemblages may represent oceanic volcanism based on the relative abundance of komatiitic flows, (Thurston and Chivers 1990, and references therein), and the scarcity of rhyolitic volcanism. Intercalations of laterally extensive sulphidic, argillite-oxide and silicate-facies iron formation suggest deep-water deposition (cf. Thurston and Chivers 1990, and references therein). The North Rim unit is in gradational, depositional and tectonic contact with the overlying Eyapamikama assemblage. The South Rim unit has been contact metamorphosed and folded into its present arcuate trace by intrusion of the crescentic 2871 million-year-old North Caribou Batholith (D.W. Davis, Royal Ontario Museum, personal communication, 1989).

The Eyapamikama assemblage overlies all units of the greenstone belt and contains 2959 million-year-old tonalite clasts (Davis, in deKemp 1987). The assemblage is in gradational depositional contact with the underlying North Rim and South Rim units, and, based upon primary structures and rock types, is interpreted as a submarine fan turbidite unit produced by erosion of older quartz-rich metasedimentary rocks. The greenstone belt is therefore a collage assembled between 2980 and 2871 Ma by a process of small-scale accretion of unrelated greenstone fragments.

Figure 5.4. Photograph of stromatolite-bearing siliciclastic material lying above the quartz arenites within the Keeyask metasedimentary assemblage.

Horseshoe Lake Greenstone Belt

This greenstone belt in the southern part of the Sachigo Subprovince (*see* Figure 5.2) was subdivided by Jensen (1987) into 4 supracrustal domains (Figure 5.5a) that we considered equivalent to the assemblages of Table 5.2. . We have subdivided the Central domain, as defined by Jensen (1987), into 2 units separated by an unconformity: the Central Horseshoe assemblage overlain by the Horseshoe assemblage. The unconformity is postulated on the basis of the stratiform carbonate and sericite alteration of the uppermost mafic flows of the underlying Central Horseshoe assemblage, and the dissimilar depositional settings of the 2 assemblages. The carbonate-sericite alteration style of mafic flows in the Central Horseshoe assemblage is similar to that observed in the Agutua Arm assemblage at the unconformity with the overlying Keeyask assemblage in the North Caribou greenstone belt (Breaks et al., in prep.). The Central Horseshoe assemblage is in gradational depositional contact with the Wapamisk assemblage of arc volcanic rocks. The tectonic contact between the south-younging Northern Horseshoe metasedimentary assemblage and the Horseshoe metasedimentary assemblage represents a "front-to-front" relationship across a shear zone.

The greenstone belt includes assemblages representing oceanic volcanism, platformal sedimentation, arc volcanism and turbiditic sedimentation (*see* Table 5.2). The

Table 5.1. Assemblages[*] of the North Caribou Lake greenstone belt.

Name	Rock Types, Primary Structures (described from base to top where known)	Interpreted Environment	Age (Ma)	References
1) Agutua Arm	basal plagioclase and hornblende-phyric massive mafic flows, rarely pillowed	subaqueous volcanism (shield)	2980	deKemp 1987
	monolithic tuff, lapilli tuff—massive, rarely thick-bedded, fragment-supported, pumice rare	proximal airfall	>2990	Davis 1990
2) Keeyask	basal 5 to 30 cm chert pebble conglomerate overlain by ~10 m supermature quartz arenite, ~5 m stromatolite quartz siltstone and ~5 m of iron formation, chert or dolomitic calcarenite	shallow-water platform	2981 detrital	Arias 1986, deKemp 1987, Davis 1990
	i) 50 m to 1.7 km thick spinifex-textured and pillowed komatiite flows and pillowed basaltic komatiites and tholeiites ii) variolitic marker unit within sequence iii) all are actinolite-rich	oceanic volcanism	<2981	Davis 1990
3) McGruer Opapimiskan Unit	i) komatiitic and tholeiitic pillowed and massive flows, selvages 3 to 5 mm, pillows ii) silicate + oxide iron formation intercalated	oceanic volcanism	?	—
Forester Unit	i) basal amphibole-phyric, pillowed and massive amygdaloidal tholeiitic and komatiitic flows, intercalated oxide and silicate iron formation ii) 60 m quartz-phyric fragmental felsic volcanic unit	oceanic volcanism	2981?	Davis 1990
South Rim Unit	i) 0.4 to 5.0 km wide pillowed + massive tholeiitic and Mg tholeiitic flows; flows are tens of metres thick, selvages are 2.5 to 25 mm thick ii) minor intermediate tuff, lapilli tuff, felsic flows and heterolithic tuff breccia; pyroclastic rocks massive to poorly bedded iii) rare oxide iron formation	oceanic volcanism	correlated with North Rim	deKemp 1987
North Rim Unit	i) 0.4 to 2.6 km wide unit of tholeiitic and komatiitic massive and pillowed mafic flows, quartz-and/or carbonate-filled amygdules ii) minor komatiitic flows high in unit iii) silicate-facies iron formation, dolomites, and felsic to intermediate tuffs	oceanic volcanism	2932 min. 2965 component	Davis 1990

Table 5.1. Assemblages* of the North Caribou Lake greenstone belt.

Name	Rock Types, Primary Structures (described from base to top where known)	Interpreted Environment	Age (Ma)	References
4) Eyapamikama	i) fining-upward assemblage—basal conglomerates, sandstone, siltstone ii) angular oligomictic volcanic breccia grades to matrix-supported polymictic cobble to boulder conglomerate; clasts include Agutua Arm, Keeyask metavolcanic and metasedimentary assemblages in arkosic-subarkosic wacke matrix; 1 cm to 2 m thick interbeds of planar to trough cross-bedded arkose, wacke iii) upper unit = slate, silt, phyllite with ball + pillow structures, rip-up clasts, slumps + flame structures	medial to distal fan turbidites	2959 min. >2870	—
5) Zeemel–Heaton	i) mature quartz wackes; quartz arenites with minor volcaniclastic units, 10 cm to 1 m thick beds of quartz-rich material of granitic provenance alternating with thick plagioclase-rich arkosic beds and rare volcanic-derived quartz-rich unit ii) less mature, poorly sorted wacke; conglomerate + mudstone, up to 60% quartz in plane-bedded mudstone; conglomerates >59% clasts, quartz-rich metasediments dominant, matrix is immature subarkosic wacke to arenite, quartz sand is 85 to 90% of matrix	correlated with Eyapikama turbidites, cannibalized platform	correlated with Keeyask	Breaks et al., in press

**The numbering format for assemblages within Table 5.1 corresponds to that used in Figure 5.2.*

Horseshoe metasedimentary assemblage and the Central Horseshoe assemblage are similar lithologically to the platformal and oceanic volcanic assemblages of the North Caribou, Upper Windigo, and Sandy Lake greenstone belts. Therefore, based on this similarity, they are tentatively assigned a Mesoarchean age of 2.9 to 3.0 Ga. The assemblages vary in structural style with homoclinal panels and folded sequences represented (*see* Figure 5.5a). The variation in structural style between assemblages and the variety of tectonic settings represented (*see* Table 5.2; Figure 5.5a) suggest that the greenstone belt represents the late assembly of diverse supracrustal fragments.

Upper Windigo Greenstone Belt

This belt (*see* Figure 5.2) is a south-younging homoclinal panel consisting largely of tholeiitic pillowed flows with subordinate medium- to fine-grained felsic pyroclastic rocks and coarse, conglomerates containing clasts of quartz arenite (Cortis et al. 1988)(Figure 5.5b). The belt is interpreted to be bounded on the south by a major dextral shear zone, the north boundary is believed to be an intrusive contact with batholiths of units 12 and 15 (*see* Bedrock Geology of Ontario, Maps 2541 to 2543, map case). The belt is correlated with the Central Horseshoe assemblage of the Horseshoe Lake greenstone belt, based on the continuity of supracrustal trends.

Favourable Lake Greenstone Belt

This greenstone belt, situated about 130 km north of Red Lake (*see* Figure 5.2), extends 160 km southeast from the Manitoba border to North Spirit Lake. The best-known part is shown in Figures 5.6a and 5.6b. The central part of the belt in the Setting Net Lake area has been the object of stratigraphic, volcanologic and geochronological work (Ayres 1970, 1972, 1974, 1977; Corfu and Ayres, in press; Corfu et al. 1985). The belt consists of 5 volcano-sedimentary cycles considered herein to be assemblages. Scale considerations have neccessitated some amalgamation of these assemblages on the Tectonic Assemblages of Ontario map (*see* Map 2575, map case). The 5 assemblages together comprise 15 formations described by Ayres (1977) and in several subsequent publications. All assemblages are bounded by positively identified or inferred thrust faults (Ayres 1988). These assemblages are described in Table 5.3.

The oldest assemblage is the Setting Net assemblage, a 2926 million-year-old komatiite-bearing unit (Corfu and Ayres, in press), possibly representing oceanic volcanism, which is succeeded by shallow-water metasedimentary rocks followed by andesitic stratovolcanic deposits suggestive of an arc environment. This assemblage is in tectonic contact with 4 successively younger arc sequences ranging from 2870 to 2725 million years in age (*see* Table 5.3).

The Setting Net assemblage is cut by the the high-level, porphyry-style, molybdenum-mineralized Setting Net Lake stock. The stock has a complex population of zircon, sphene and monazite, interpreted by Corfu and Ayres (in press) to indicate:

1. entrainment of xenocrystic zircons with ages of 2743, 2720 and 2715 Ma

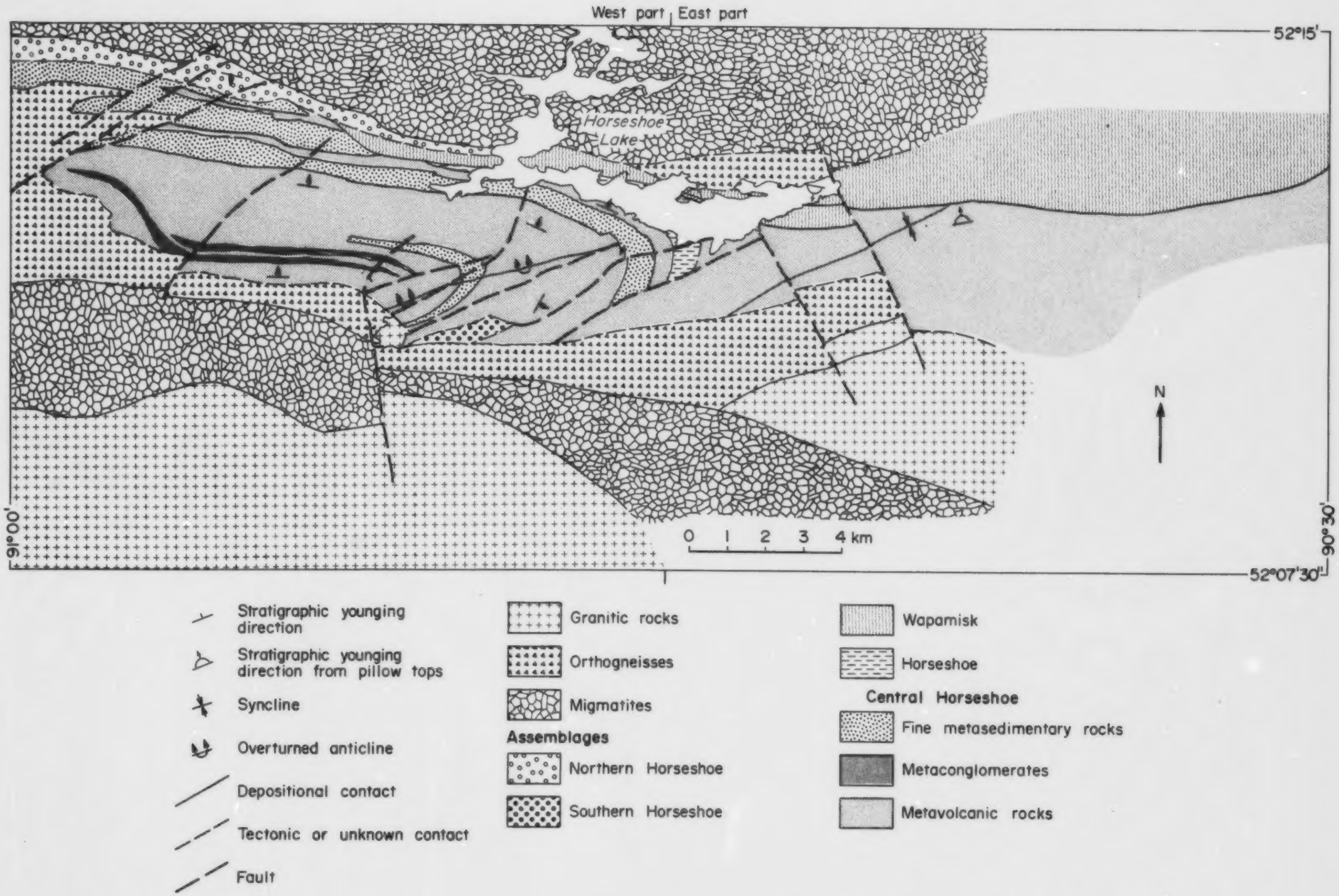

West part
East part
52°15'
Horseshoe Lake
N
0 1 2 3 4 km
91°00'
90°30'
52°07'30"
Stratigraphic younging direction
Stratigraphic younging direction from pillow tops
Syncline
Overturned anticline
Depositional contact
Tectonic or unknown contact
Fault
Granitic rocks
Orthogneisses
Migmatites
Assemblages
Northern Horseshoe
Southern Horseshoe
Wapamisk
Horseshoe
Central Horseshoe
Fine metasedimentary rocks
Metaconglomerates
Metavolcanic rocks

Figure 5.5a. Sketch map of the geology of the Horseshoe Lake greenstone belt (*modified from* Jensen 1988). Assemblages based upon the analysis in this chapter are superimposed on the subdivisions described by Jensen (1988).

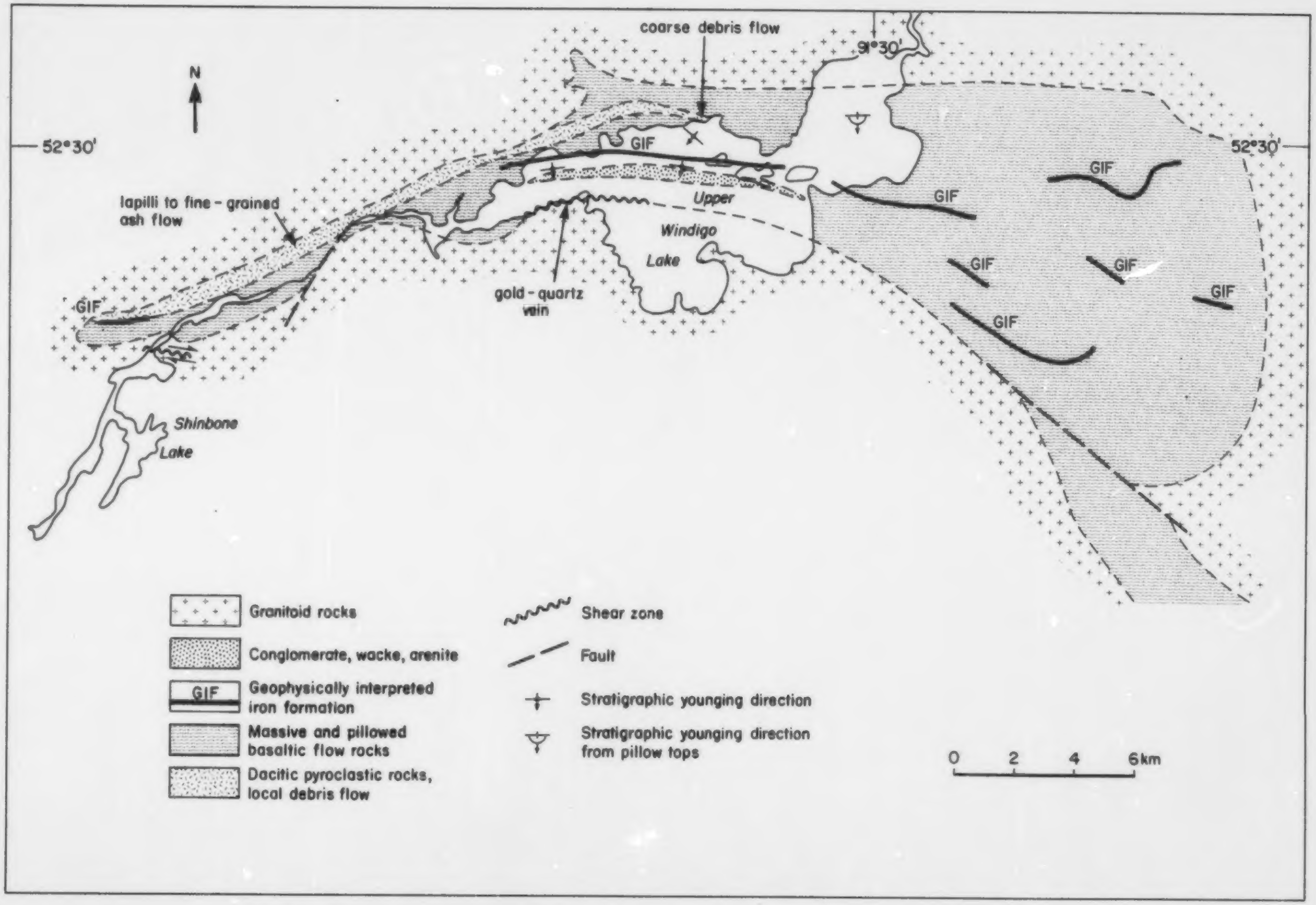

coarse debris flow
91°30'
N
52°30'
52°30'
GIF
GIF
GIF
GIF
GIF
GIF
GIF
GIF
lapilli to fine - grained ash flow
Upper
Windigo
Lake
gold - quartz vein
Shinbone
Lake
Granitoid rocks
Conglomerate, wacke, arenite
GIF
Geophysically interpreted iron formation
Massive and pillowed basaltic flow rocks
Dacitic pyroclastic rocks, local debris flow
Shear zone
Fault
Stratigraphic younging direction
Stratigraphic younging direction from pillow tops
0
2
4
6km

Figure 5.5b. Sketch map of the Upper Windigo greenstone belt *from* Cortis et al. (1988).

Table 5.2. Assemblages* of the Horseshoe Lake greenstone belt.

Name	Rock Type, Primary Structures	Contacts	Environment
6) Central Horseshoe	i) Mg and Fe tholeiitic massive + pillowed flows, upward increasing proportion of intercalated clastic and lesser chemical metasediments ii) clastic sediments progress upward from pebbly wackes to mature quartz arenite + quartzofeldspathic arenite; conglomerates contain clasts of calc-alkalic rhyolite, cherty metasediments, fuchsitic cherty siltstone + mafic metavolcanics iii) chemical metasediments — oxide + sulphide iron formation, marble iv) uppermost unit (400 m thick) — limestone, chert, wacke, and quartz arenite	tectonic contact with Southern Horseshoe depositional contact with Wapamisk tectonic(?) contact with gneisses tectonic(?) contact with Northern Horseshoe unconformity(?) with Horseshoe	oceanic volcanism
7) Wapamisk	i) calc-alkalic basalt and andesite pillowed flows ii) massive flows and fine-grained pyroclastic rocks iii) intercalations of iron formation and graphitic units	depositional contact with Central Horseshoe tectonic(?) contact with Northern Horseshoe	arc volcanism distal facies
8) Northern Horseshoe	quartz-rich arenites, wackes, oxide + sulphide facies iron formation with minor tholeiitic pillowed flows	tectonic contact with migmatites to north tectonic contact with Central Horseshoe and Wapamisk	resedimented metasediments
9) Southern Horseshoe	i) conglomerates with arkosic matrix, wacke and siltstones ii) 10 to 50 m thick, fining-upward cycle with basalt monomictic matrix-supported conglomerate followed by wacke and siltone; clasts largely wackes	tectonic contact with migmatites unknown type of contact with Central Horseshoe (shear or unconformity)	resedimented metasediments
10) Horseshoe	limestone, chert, wacke and quartz arenite, with graded and cross-bedded units; sequence is 400 m thick	unconformable(?) contact with Central Horseshoe tectonic contact with Wapamisk	platformal sedimentation

** The numbering format for assemblages within Table 5.2 corresponds to that used in Figure 5.2.*

2. magmatic crystallization indicated by an upper intercept zircon age of 2710 to 2712 Ma, and an emplacement age of 2708 Ma indicated by titanite data
3. subsolidus crystallization and hydrothermal alteration shown by a sequence of titanite and rutile ages ranging from 2706 to 2657 Ma

The 2870 Ma date for the South Trout assemblage is based on a baddeleyite age of a subvolcanic diorite sill (Corfu and Ayres, in press). This age is interpreted as the age of volcanism, but may represent a minimum age for the assemblage. The alluvial fan conglomerates of the 2725 million-year-old North Trout assemblage include clasts of 2959 million-year-old granodiorite, clearly suggesting that erosion of older sial is involved in the genesis of 2.7 billion-year-old metasedimentary rocks in the southern part of the Sachigo Subprovince (Corfu and Ayres, in press).

The volcanism of the Northwind and North Trout assemblages was synchronous with crystallization of the earliest widespread phase of the North Trout Lake batholith on the north side of the greenstone belt, at 2732 Ma (Corfu and Ayres, in press). The absence of volcanic units less than 2725 million years old correlative with the younger phases of the batholith suggest that if they existed, they have probably been eroded, as only midcrustal levels (amphibolite facies) of the greenstone belt are present. The 2725 million-year-old volcanism was followed by thrusting, folding around east- to southeast-striking axes, metamorphism, and intrusion of the 2711 million-year-old phases of the North Trout Lake batholith (Corfu and Ayres, in press; Corfu and Davis, this volume). Metamorphism culminated between the age of the youngest widespread phase of the North Trout Lake batholith (2711 Ma) and the age of titanite (2679 Ma) in the amphibolite-facies granodiorite clasts within conglomerates of the North Trout assemblage.

North Spirit Lake Greenstone Belt

The North Spirit Lake greenstone belt lies north-northeast of Red Lake, within the southwestern part of the Sachigo Subprovince (*see* Figure 5.2). The greenstone belt has been mapped by Bateman (1939) and Wood (1977, 1980a, 1980b; Corfu and Wood 1986). Lithologic descriptions are available in Table 5.4, and the distribution of assemblages is portrayed on Figure 5.7. Changes in the stratigraphic nomenclature of Corfu and Wood (1986) are incorporated in Table 5.4 and Figure 5.7.

The North Spirit assemblage comprises 3023 million-year-old arc volcanic rocks, the oldest supracrustal rocks in the Superior Province (Corfu and Wood 1986). The assemblage is isolated from the remainder of the greenstone belt by the North Spirit shear zone. The Disrupted assemblage, a migmatized arc volcanic sequence, forms the eastern margin of the belt, south of the North Spirit shear zone. This assemblage is dated at 2900 to 2950 Ma (Corfu and Wood 1986); however, there is probably an inherited component in the zircon population. The Disrupted assemblage is capped by extensive iron formation (*see* Figure 5.7) overlain by the platformal Nemakwis assemblage. Quartz arenites near the tectonically and metamorphically complex contact zone between the Disrupted assemblage and the Nemakwis assemblage contain detrital 2986 million-year-old zircons. However, the quartz arenites cannot be unequivocally correlated with the platformal Nemakwis assemblage; although it is unlikely they would be part of the arc-related units of the Disrupted assemblage.

The upper contact of the Nemakwis assemblage is a regional-scale disconformity to unconformity (Wood 1980a, 1980b) above which lies the Makataiamik assemblage of uncertain age. The Makataiamik assemblage is a fining-upward sequence of clastic metasedimentary rocks, succeeded by marble and iron formation, and overlain by an uppermost unit of carbonatized mafic flows and serpentinite. The assemblage youngs to the east along the western margin of the belt. Similar rock types in the area of Hewitt Lake are correlated with the Makataiamik assemblage.

The Hewitt Lake area is bounded by major shear zones to the northeast and southwest and is structurally complex. South of Hewitt Lake are homoclinal panels of the Hewitt assemblage, whereas north of Hewitt Lake are complexly folded rocks correlated with the Makataiamik assemblage. Therefore, the area north of Hewitt Lake area is considered to be a tectonic window through the Hewitt assemblage into the underlying Makataiamik assemblage (*see* Figure 5.7).

The metasedimentary rocks of the Makataiamik assemblage and the Nemakwis assemblage are relatively quartz-rich compared to metasedimentary rocks of most Archean greenstone belts (Donaldson and Jackson 1965; Donaldson and Ojakangas 1977; Wood 1980a). The geochronological data indicate that the quartz-rich metasediments are derived from erosion of an approximately 2.9 billion-year-old sialic protolith during both the platformal sedimentation of the Nemakwis assemblage and the alluvial fan stage of the Makataiamik assemblage.

The approximately 2.7 billion-year-old Hewitt assemblage arc volcanic rocks are similar lithologically to arc sequences elsewhere in the Superior Province, save for the greater thickness and possibly the deeper-water paleoenvironmental setting of arc-derived metasediments forming the uppermost part of the assemblage (Wood 1980a).

The Bijou Point complex (Smith and Longstaffe 1974) is geochemically comparable to shoshonitic assemblages forming late unconformable sequences in the Abitibi (Cooke and Moorhouse 1969) and Wabigoon (Blackburn 1982) subprovinces (*see* Blackburn et al. and Jackson and Fyon, this volume). The tectonic position of the sequence, spatially associated with the North Spirit shear zone and the Bear Head fault zone (Osmani and Stott 1988) is also comparable with many of the other late unconformable assemblages elsewhere in Superior Province (Wyman and Kerrich 1988).

Wunnummin Lake Greenstone Belt

The Wunnummin Lake greenstone belt lies north of Pickle Lake in the southeastern part of the Sachigo Subprovince (*see* Figure 5.2). Mapping in the area has included work by Prest (1942), Thurston et al. (1979), Stott and Janes (1985), and Thurston et al. (1987). The belt consists of 4 assemblages (Figure 5.8), described in Table 5.5: the Wunnummin assemblage, the Peeagwon assemblage, the Big North assemblage and the Central Wunnummin assemblage. The greenstone belt as a whole has been metamorphosed to mid-greenschist facies in the interior of the belt, and to amphibolite facies at the margins.

The 2 assemblages south of the Stull Lake–Wunnummin Lake fault zone (SLWLFZ) are platformal, based on the occurrence of komatiites and intercalated quartz-rich metasediments (Stott and Janes 1985; Thurston et al. 1987). Detrital zircons in the Wunnummin assemblage indicate a source age of 2965 Ma, comparable with the provenance age of most other Sachigo Subprovince platformal greenstones. The Peeagwon and Wunnummin assemblages have comparable rock types and both lie south of the SLWLFZ. They are tentatively correlated on this basis. The complex trace of stratigraphic marker units (Thurston et al. 1979) in the Wunnummin assemblage near Maria Lake suggests that the Wunnummin assemblage is intensely deformed relative to other greenstones in the belt. The Big North assemblage represents long, linear panels of arctype greenstones; as an extension of the Big Trout Lake–Swan Lake greenstone belt, we suggest the Big North assemblage may be of comparable age (2.83 Ga).

The Central Wunnummin assemblage is a unit of alluvial-fluvial metasediments in disconformable contact with the underlying Wunnummin assemblage (Prest 1942). The position of the former assemblage along the SLWLFZ, its disconformable relationship to the Wunnummin assemblage, and the inferred tectonic environment suggest that the Central Wunnummin assemblage represents late-stage rocks laid down unconformably on the Wunnummin assemblage. Given the fact that the Central Wunnummin assemblage lies along the SLWLFZ, it may be similar in age to the Oxford Lake Group which has been dated at 2709 Ma (cf. MEM 1987). The SLWLFZ may represent a major tectonic boundary between the arc-affiliated 2.8 billion-year-old greenstones to the north and the platformal 2.9 billion-year-old greenstones to the south.

Muskrat Dam Lake Greenstone Belt

The Muskrat Dam Lake greenstone belt lies north-northeast of Red Lake, in the Sachigo Subprovince (*see* Figure 5.2). The belt has been mapped by Ayres (1969) on a reconnaissance scale with subsequent work carried out by Thurston et al. (1987). The belt is complexly folded, and only 3 samples have been radiometrically dated (*see* Table 5.6), therefore stratigraphic and structural interpretations are tentative.

We divide the belt into 3 major assemblages: a Nekence assemblage consisting of 2.9 billion-year-old platformal units, a Muskrat Dam assemblage consisting of 2.7 billion-year-old arc volcanic rocks, and a 2.9 billion-year-old (?) Rottenfish assemblage (Figure 5.9, Table 5.6). The Nekence assemblage includes komatiitic and tholeiitic flows at Nekence Lake, and overlying iron formation and quartz arenites. This sequence was considered part of a "lower metavolcanic formation" by Ayres (1969). Ayres correlated the unit with similar rocks at the west end of the greenstone belt; however, exposures in that area have not been re-examined and the suggested correlation is uncertain. The contact between the Nekence assemblage and the Muskrat Dam assemblage is of uncertain nature. The shear zone separating the Keeyask assemblage from younger units within the North Caribou greenstone belt may extend northwest to form the contact between the major assemblages within the Muskrat Dam Lake greenstone belt (cf. Tectonic Assemblages of Ontario, Maps 2575 to 2577, map case).

The Muskrat Dam assemblage includes rhyolitic units in 2 stratigraphic positions which have been dated. Both samples yielded an age of 2734 Ma (D.W. Davis, Royal Ontario Museum, personal communication, 1989). Komatiitic flows have been observed (Thurston et al. 1987) along the Severn River within the gabbroic units. The upper metasedimentary unit described by Ayres (1969) (*see* Figure 5.9) includes, at the west end of Sandhill Crane Island, a unique conglomerate with a carbonate matrix and a clast population as follows: matrix 41 to 61%, gabbro 0 to 26%, mafic metavolcanic rock 10 to 36%, felsic metavolcanic rock 0 to 3%, metasediments 0 to 6%, chert 5 to 16 %, iron formation 0 to 5%, quartz arenite 0 to 2%, and vein quartz 0 to 2%. Some clasts are surrounded by delicately layered carbonate (Figure 5.10), tentatively identified as stromatolitic in origin (Thurston et al. 1987). The presence of rhyolitic units of similar age at 2 stratigraphic positions, and the presence of possibly non-arc komatiitic volcanic units within the Muskrat Dam assemblage is indicative of unrecognized (thrust-based?) stratigraphic repetition. The presence of quartz arenite clasts, which are probably 2.9 billion years of age, in the 2734 million-year-old Muskrat Dam assemblage means that it was, in part, derived from erosion of 2.9 billion-year-old Sachigo Subprovince material prior to the emplacement of the rhyolitic part of the assemblage (2734 Ma) (D.W. Davis, Royal Ontario Museum, personal communication, 1990).

The Rottenfish assemblage is separated from the main Muskrat Dam Lake greenstone belt by the Rottenfish Shear Zone (Osmani and Stott 1988). The trend of the belt, and its link via a xenolith train to the Sandy Lake greenstone belt,

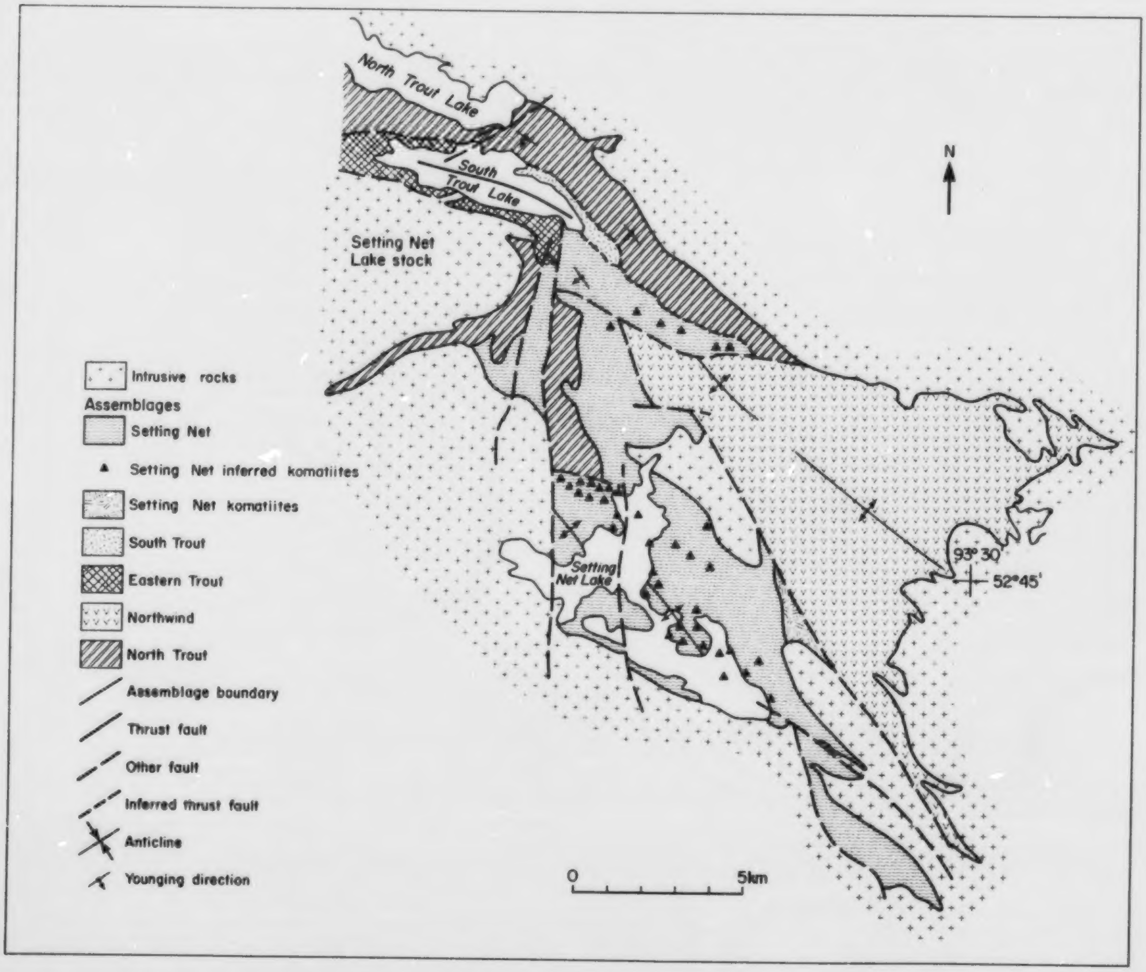

North Trout Lake
South Trout Lake
Setting Net Lake stock
Setting Net Lake
N
93°30'
52°45'
0
5km
Intrusive rocks
Assemblages
Setting Net
Setting Net inferred komatiites
Setting Net komatiites
South Trout
Eastern Trout
Northwind
North Trout
Assemblage boundary
Thrust fault
Other fault
Inferred thrust fault
Anticline
Younging direction

Figure 5.6a. Assemblages of the Favourable Lake greenstone belt based on the stratigraphic nomenclature of Ayres (1977).

favours correlation of the Rottenfish assemblage with the 2.9 billion-year-old North Sandy Lake assemblage of arc metavolcanic rocks at Sandy Lake.

Sandy Lake Greenstone Belt

The Sandy Lake greenstone belt lies north of Red Lake in the southwestern part of the Sachigo Subprovince (*see* Figure 5.2). It has been mapped by Satterly (1939) with additional work by Thurston et al. (1987) and Cortis (1988). The belt consists of 4 metavolcanic-metasedimentary assemblages separated by branches of the Sandy Lake shear zone (Osmani and Stott 1988) (Figure 5.11, Table 5.7). In general, the greenstone belt is a south-younging arc sequence overlain by quartz-arenite–bearing submarine-fan and wrench-basin assemblages. The 2945 million-year-old North Sandy assemblage is very well preserved with silicified pumice in ash-flows indicative of vapour-phase recrystallization immediately following the eruption, which suggests that the felsic part of the assemblage was deposited subaerially (Thurston et al. 1987). The North Sandy assemblage is approximately correlative with the 2926 million-year-old Setting Net assemblage arc sequence at Favourable Lake.

The quartz-rich metasedimentary rocks of the Keewaywin and Sandborn assemblages represent the erosion of pre-existing platform sequences (Thurston et al. 1987; Cortis 1988). The age of the detrital zircons (dated at 2.9 to 3.0 Ga) is similar to the age of detrital zircons in quartz-rich metasediments throughout the southern part of the Sachigo Subprovince. The fact that the Keewaywin sequence sits on a tectonized contact above the 2945 million-year-old North Sandy assemblage suggests that no great time interval exists between the deposition of the arc assemblage and the submarine-fan assemblage.

The West Arm assemblage may represent an oceanic environment on the basis of the abundance of chert clasts in the conglomerates. The presence of abundant quartz arenite clasts suggests that the sequence may be similar to the other quartz-arenite–bearing assemblages derived from 2.9 billion-year-old sialic precursors.

Lingman Lake Greenstone Belt

This greenstone belt (*see* Figure 5.2) lies 325 km north of Red Lake and consists of a single complexly folded assemblage (Wilson 1987) dominated by mafic tholeiitic massive and pillowed flows with minor felsic tuff, crystal tuff and arenaceous metasedimentary rocks (Figure 5.12). The assemblage is in intrusive contact with marginal granitic rocks and is disrupted by a major east-striking dextral shear zone. The basal unit consists of massive and pillowed tholeiitic flows, commonly less than 10 m thick, separated by interflow tuff and arenite units usually less than 1 m thick. Pillows up to 1.5 by 2 m display "pillow-shelf" structures (Budkewitsch and Robin 1989), and small variolites in cores and rims. Both feldspar-phyric flows and hornblende-phyric intermediate to mafic tuff occur toward the top of the assemblage. Felsic metavolcanic rocks and metasediments, separated by mafic flows, form the upper part of the belt, defining 3 to 5 discrete intervals of pyroclastic and sedimentary activity. Felsic metavolcanic rocks consist of flows and thickly bedded tuff and crystal tuff interbedded with arenite, conglomerate, wacke and mudstone (Wilson 1987). Metasedimentary units of the belt include a unit of arenite, a wacke-mudstone association, and conglomerate with a matrix consisting of up to 25% quartz granules measuring 1 mm in diameter. The conglomeratic rocks include clasts of vein quartz, chert, quartz arenite, and metavolcanic rock in a

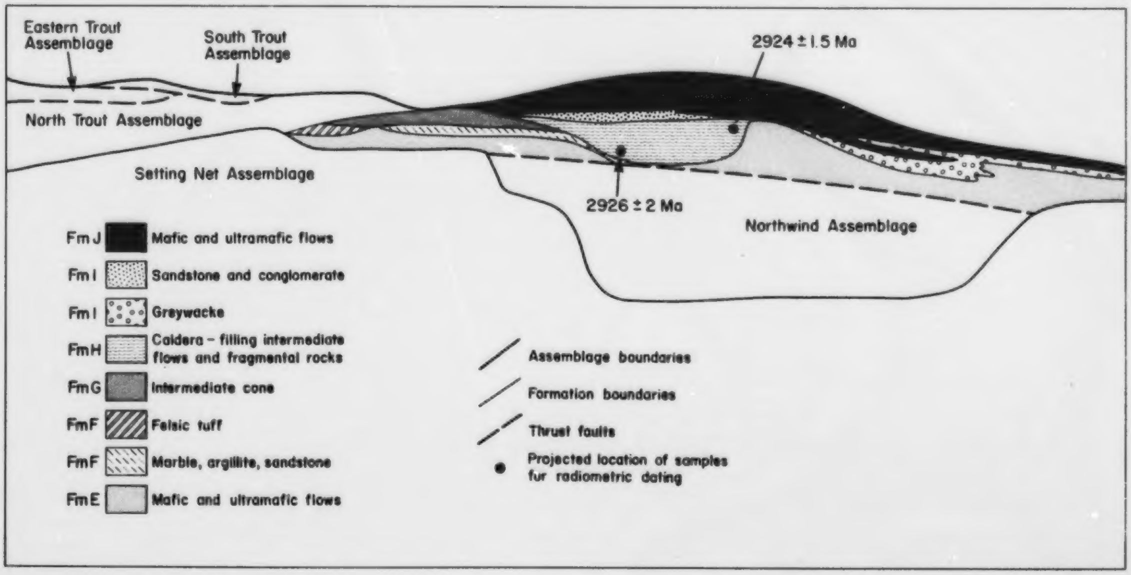

Eastern Trout Assemblage
South Trout Assemblage
2924 ± 1.5 Ma
North Trout Assemblage
Setting Net Assemblage
2926 ± 2 Ma
Northwind Assemblage
FmJ ■ Mafic and ultramafic flows
FmI ▒ Sandstone and conglomerate
FmI ∴ Greywacke
FmH ▒ Caldera - filling intermediate flows and fragmental rocks
FmG ▨ Intermediate cone
FmF ▨ Felsic tuff
FmF ▨ Marble, argillite, sandstone
FmE □ Mafic and ultramafic flows
Assemblage boundaries
Formation boundaries
Thrust faults
● Projected location of samples fur radiometric dating

Figure 5.6b. Stratigraphic cross section of part of the Favourable Lake greenstone belt with major formations described by Ayres (1977).

quartz-rich matrix. These conglomerates are petrographically similar to the resedimented conglomerates of the Eyapamikama assemblage. Quartz gabbro dikes cut the supracrustal rocks of the belt and the surrounding granitoid units. Based on structural trends and rock types, this greenstone belt is correlated with the 2.9 billion-year-old greenstones of the North Caribou terrane (cf. Davis in deKemp 1987).

Pierce–Ponask–Sachigo Greenstone Belt

The Ponask Lake shear zone (Thurston et al. 1987; Osmani and Stott 1988) separates this belt (*see* Figure 5.2) into a Sachigo assemblage and a Pierce assemblage. The assemblages are characterized by distinctive rock types (Table 5.8). The south-younging Pierce assemblage (Figure 5.13) comprises metavolcanic units including ash flows with arc affinities. The Pierce assemblage is correlated with the Island Lake area of Manitoba based on the following: 1) its position south of the Ponask Lake shear zone, 2) the arc affinity of the metavolcanic rocks, and, 3) the continuity (based upon aeromagnetic interpretation using proprietary magnetic data) of the granitic rocks bordering the belt with granitic rocks bordering the Island Lake greenstone belt. These granitic rocks display a low to intermediate magnetic intensity and a striped magnetic pattern. The Sachigo assemblage is characterized by 1) quartz-rich metasediments, 2) a spatial association with the highly magnetic plutons of the Munro Lake area, and 3) continuity with the easterly trending magnetic lineaments characterizing the Munro Lake belt (MEM 1987). Therefore, the Sachigo

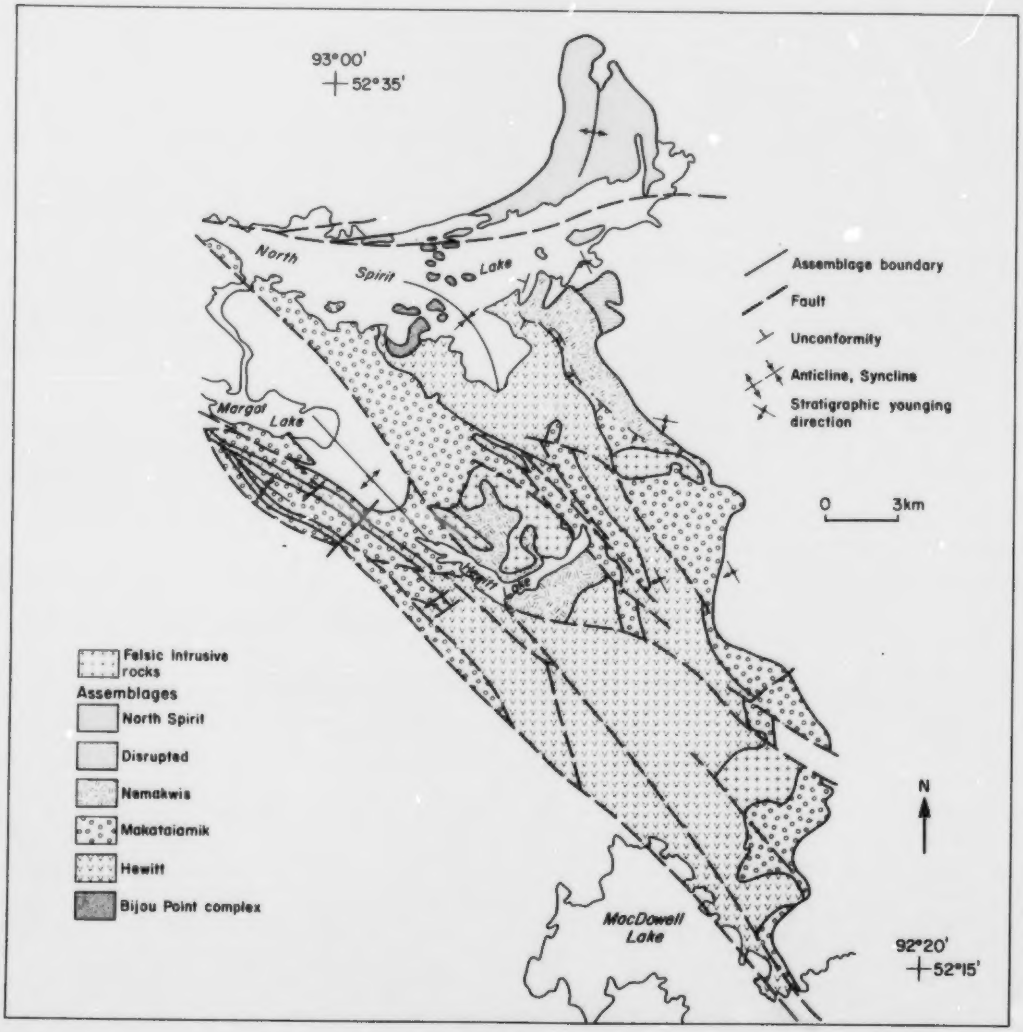

93°00'
52°35'
North Spirit Lake
Margot Lake
MacDowell Lake
Assemblage boundary
Fault
Unconformity
Anticline, Syncline
Stratigraphic younging direction
0 3km
N
Felsic intrusive rocks
Assemblages
North Spirit
Disrupted
Nemakwis
Makataiamik
Hewitt
Bijou Point complex
92°20'
52°15'

Figure 5.7. Assemblages of the North Spirit Lake greenstone belt derived from mapping carried out by Wood (1980a, 1980b) and the geochronological study done by Corfu and Wood (1986).

assemblage may be correlative with the 2760 million-year-old Munro Lake greenstones.

Big Trout Lake–Swan Lake Greenstone Belt

STULL LAKE PORTION

This greenstone belt is the northernmost major greenstone belt in Ontario. For purposes of description, it is subdivided into 2 parts. The Stull Lake portion of the Big Trout Lake–Swan Lake greenstone belt is the western end of the belt within Ontario, lying along the Ontario–Manitoba border (*see* Figure 5.2). The western extension of the belt into Manitoba includes greenstones at Gods Lake and Oxford Lake. Previous mapping within the belt includes that of Satterly (1938), Bennett and Riley (1969) and Thurston et al. (1987). The Stull Lake portion is divided into 3 assemblages (Figure 5.14) bounded mainly by the Stull Lake–Wunnummin Lake fault zone of Osmani and Stott (1988).

The Stull assemblage (Table 5.9) is an arc assemblage correlated with the 2.8 billion-year-old Hayes River Group (MEM 1987). The suggested correlation is made on the basis of structural trends, presence of plagioclase-phyric basaltic units and gabbroic intrusions in both locations, and the nearly continuous nature of the "arc-type" sequences between Oxford Lake and Stull Lake.

The Rapson assemblage is considered (Satterly 1938), on the basis of rock types to be a "Timiskaming-type" sequence. Given the relative continuity of this type of unit between Oxford Lake and Stull Lake, the assemblage is tentatively correlated with the 2706 million-year-old Oxford Lake Group (MEM 1987).

The Richardson Arm assemblage consists of hornblende-phyric massive and pillowed mafic flows, which, on the basis of structural trends and rock types, is correlated with the 2.8 billion-year-old units of the Hayes River Group.

BIG TROUT LAKE–SWAN LAKE PORTION

Previous mapping in this portion of the greenstone belt includes Hudec (1964), Bennett and Riley (1969), and Thurston et al. (1979). The western end of the belt consists of 2 arms, each about 3 km wide, which extend from the Hanson River to the eastern end of Big Trout Lake. The 2 arms are disrupted by the Stull Lake–Wunnummin Lake fault zone (SLWLFZ), but younging data support the interpretation that these 2 arms form the limbs of an anticlinorium, which is cored by granitic batholiths (Bennett and Riley 1969).

This portion of the belt consists of about 75% mafic metavolcanic rocks, 10% felsic to intermediate metavolcanic rocks, 10% metasediments and 5% mafic intrusive rocks. The Big Trout Lake–Swan Lake portion of the greenstone belt is poorly known and is considered to be one unsubdivided assemblage on the Tectonic Assemblages of Ontario map (*see* Map 2575, map case).

The mafic metavolcanic rocks include pillowed and massive flows and layered amphibolites containing plagioclase-, hornblende- and epidote-rich layers. These amphibolites are especially common between the Severn River and Big Trout Lake. Primary structures are well preserved on eastern Big Trout Lake, including variolitic and flow-banded units, primary clinopyroxene and calcic plagioclase. Mafic flows in the Big Trout Lake area are commonly epidotized, particularly south of Big Trout Lake. Mafic flows include thick (greater than 100 m), possibly subaerial, units lacking pillows (Thurston et al. 1979); but subaqueous pillowed to massive flows predominate. Dacitic flows and tuffs occur at Swan Lake and in the eastern part of Big Trout Lake. Coarse fragmental rocks occur to the west of the north shore of Big Trout Lake, along the northern flank of the Big Trout Lake–Swan Lake greenstone belt, and along the south flank of the southern arm of the western part of the greenstone belt.

Metasedimentary rocks are scarce; wacke and conglomerate are dominant, forming 2 major units between the Severn River and Big Trout Lake, and on Bearskin Lake. Conglomerates on eastern Big Trout Lake have a clast population as follows: 60% felsic metavolcanic rock, 40% mafic metavolcanic rock, and less than 1% chert clasts. The matrix makes up about 40% of the unit. Minor units of "Timiskaming" type conglomerates, containing clasts of jasper and granite, occur near the Sachigo River (Meen 1938), south of Corking Lake.

The supracrustal rocks of the greenstone belt are cut by the Big Trout Lake intrusive complex (Macdonald and Cherry 1988), a pretectonic to syntectonic layered sill with an ultramafic lower part and a fractionated mafic upper part (*see* Figure 5.8). The Big Trout Lake intrusive complex is mineralized with chromite, and platinum group element concentrations are currently being explored (Janes et al. 1990). An ophitic-textured gabbroic stock about 25 km² in area occurs about 26 km west of Swan Lake. A small pluton of serpentinite occurs at Misikeyask Lake, spatially associated with the SLWLFZ along the south boundary of the belt.

The Kino Lake greenstone belt (Hudec 1964) is that part of the Big Trout Lake greenstone belt south of the SLWLFZ. The belt is composed of massive, coarse, mafic to intermediate flows with intercalated wackes and "impure" quartzites. A small peridotite body is spatially associated with the sheared northern contact of the belt. The Kino Lake greenstone belt may represent an oceanic depositional environment, as it lacks rhyolite, and is therefore correlated with the Western Assemblage of the Pierce–Ponask–Sachigo greenstone belt.

Blackbear River–Ellard Lake Greenstone Belt

This greenstone belt has been traced for 130 km from Rorke Lake on the Ontario–Manitoba border, eastward to the Blackbear River (*see* Figure 5.2). The belt is composed largely of medium metamorphic grade mafic pillowed flows, with minor amphibolite occurring near the Sachigo

Table 5.3. Assemblages* of the Favourable Lake greenstone belt.

Name	Rock Types, Primary Structures (stratigraphy described from base to top)	Interpreted Environment	Age (Ma)	References
11) Setting Net	– Cycle 2 of Ayres (1977) entire assemblage 2.3 km thick; both contacts are thrust faults		2926	Corfu and Ayres, in press Ayres 1977
	Fm E komatiitic and tholeiitic pillowed and massive flows, rare variolitic textures, 300 to 2200 m thick; komatiite flows 1 to 5 m thick, pillowed and spinifex textured, form 640 m thick unit overlain by tholeiites	oceanic (?) sequence		
	Fm F stromatolitic (?) marble, chert, siltstone, sandstone, ferruginous chert, form 0 to 60 m thick unit	shield volcano		
	Fm G andesite and dacite pyroclastic rocks, lenticular, 0 to 630 m thick unit; vulcanian breccias, quartz-rich, thin and thick bedded	subaerial, vent to distal facies, caldera structure		
	Fm H dacite pyroclastic rocks, flows, vulcanian breccia, accretionary lapilli, form 0 to 1370 m thick unit	caldera fill		
	Fm I proximal sandstones, conglomerates, thick graded beds succeeded by thin bedded chert-argillite-iron formation	top of caldera-fill sequence		
Setting Net (?)	Fm J mafic pillowed flows, minor pyroclastic rocks	subaqueous volcanism		
12) South Trout	– Cycle 5 of Ayres (1977) entire assemblage 0 to 580 m thick, both contacts are thrust faults:		2870	Corfu and Ayres, in press
	Fm P–1 wacke-argillite, rare slate, chert iron formation; thin to medium bedded, no grading; 0 to 215 m thick unit	marine		
	Fm P–2 andesite to dacite flows, pyroclastic rocks with intercalated iron formation, amygdules, vulcanian breccia; 0 to 365 m thick unit	shallow water to subaerial		
13) Eastern Trout	– Cycle 4 of Ayres (1977) entire assemblage 0 to 600 m thick; lower contact is a thrust fault:		2858	
	Fm M mafic flows with rare mafic and felsic pyroclastic rocks; form 0 to 480 m thick unit; pillows absent at west end of belt	partially subaerial arc sequence		Corfu and Ayres, in pre

Table 5.3. Assemblages* of the Favourable Lake greenstone belt.

Name	Rock Types, Primary Structures (stratigraphy described from base to top)	Interpreted Environment	Age (Ma)	References
	Fm N felsic tuff with lesser intermediate tuff—all reworked; form 0 to 120 m thick unit; thin to medium bedded			
14) Northwind	– Cycle 1 of Ayres (1977) 4850 m thick thrust at base:		2734	Ayres 1977
	Fm A intermediate flows, minor pyroclastic units, minor felsic flows and pyroclastic units, iron formation; interlayering of flows and pyroclastic units; heterolithic pyroclastic units, vulcanian breccia; 3000 + m thick unit	shield volcano(?)		
	Fm B–1 ferruginous chert, minor marble, slate, sandstone, tuff, form 0 to 260 m thick unit	shallow submarine		
	Fm B–2 mafic flows, minor intermediate to felsic tuff; pillowed; 200 to 365 m thick unit	proximal submarine		
	Fm C felsic flows, massive to locally brecciated; 0 to 670 m thick unit	proximal to distal submarine		
	Fm D–1 iron formation, minor intermediate to mafic tuff, argillite; form 150 m thick unit			
	Fm D–2 argillite-siltstone, minor sandstone, conglomerate; thin to thick bedded, graded; 90 to 140 m thick unit			
15) North Trout	– Cycle 3 of Ayres (1977) entire assemblage is 2 to 5 km thick:		2725	Corfu and Ayres, in press
	Fm J mafic pillowed flows, minor mafic to intermediate tuff, lapilli tuff, sandstone, slate, ferruginous chert; form 0 to 1450 m thick unit	deepening basin		
	Fm K–1 conglomerate, wacke, calc-silicates; medium to thick bedded with grading; 0 to 425 m thick unit	subaerial provenance alluvial fan		Gordanier 1982
	Fm K–2 wacke with minor felsic tuff and flows, ferruginous chert, marble; thin to medium bedded with grading; carbonate alteration			

**The numbering format for assemblages within Table 5.3 corresponds to that used in Figure 5.2.*

Table 5.4. Assemblages* of the North Spirit Lake greenstone belt.

Name	Rock Types, Primary Structures (stratigraphy described from base to top)	Interpreted Environment	Age (Ma)	References
16) North Spirit	mafic pillowed and massive flows, amygdaloidal flows, felsic tuff flows, tuff, lapilli tuff, tuff breccia, oxide-facies iron formation	arc volcanism	3023 crystallization	Corfu and Wood 1986
17) Disrupted	basal massive mafic flows, intermediate to felsic flows and pyroclastic rocks, minor quartz arenites, oxide-facies iron formation (upper unit)	arc sequence	2900 to 2950 crystallization	Corfu and Wood 1986
18) Nemakwis Lake	lower contact = iron formation of Disrupted assemblage; overlain by quartz arenites, oxide iron formation and spinifex-textured and pillowed komatiitic flows and ultramafic metasediments, minor intermediate tuffs; cross-beds in arenites; ultramafic metasediments medium to thin bedded; total thickness ~ 600 m; upper contact = unconformity	platform	2986 detrital	Corfu and Wood 1986 Wood 1977
19) Makataiamik	– Lower unit of Corfu and Wood (1986) lower contact = unconformity; 1900 m thick fining-upward unit; 460 m framework-supported conglomerates, scour structures, quartz arenite interbeds; gradational contact with 1400 m thickness of sandstones, succeeded by iron formation; conglomerates contain clasts of felsic, mafic and weathered ultramafic metavolcanics in a matrix of poorly sorted wacke; interbedded sandstones are graded, massive, or equigranular with dish structures and parallel laminae of bed tops; minor lithic arkose, quartz arenite and subarkose are intercalated with conglomerates; upper sandstone unit includes Bouma A, B, C, A–D, E divisions; conglomerate clasts include 2975 to 3001 Ma tonalites; conglomerate succeeded laterally by carbonatized mafic flows then marble, iron formation and felsic tuff, lapilli tuff; upper contact = tectonic	alluvial fan grading up to braided fluvial to lacustrine	<2900 to 2950 >2735	—
20) Hewitt	– Upper unit of Corfu and Wood (1986) lower contact = tectonic; basal basaltic and dacitic flows and coarse- to fine-grained pyroclastic rocks overlain by thick bedded granular arkose sand conglomerates, with minor iron formation, marble, chert	arc volcanics deep water turbidites	2735 crystallization	—
21) Bijou Point complex	intermediate to felsic tuffs, derived conglomerates and sandstones cut by 2731 Ma subvolcanic quartz diorite to granodiorite; conglomerates have sandstone interbeds; volcanic units have shoshonitic geochemistry	late alkalic volcanism fluvial sedimentation	2731	Smith and Longstaffe 1974

**The numbering format for assemblages within Table 5.4 corresponds to that used in Figure 5.2.*

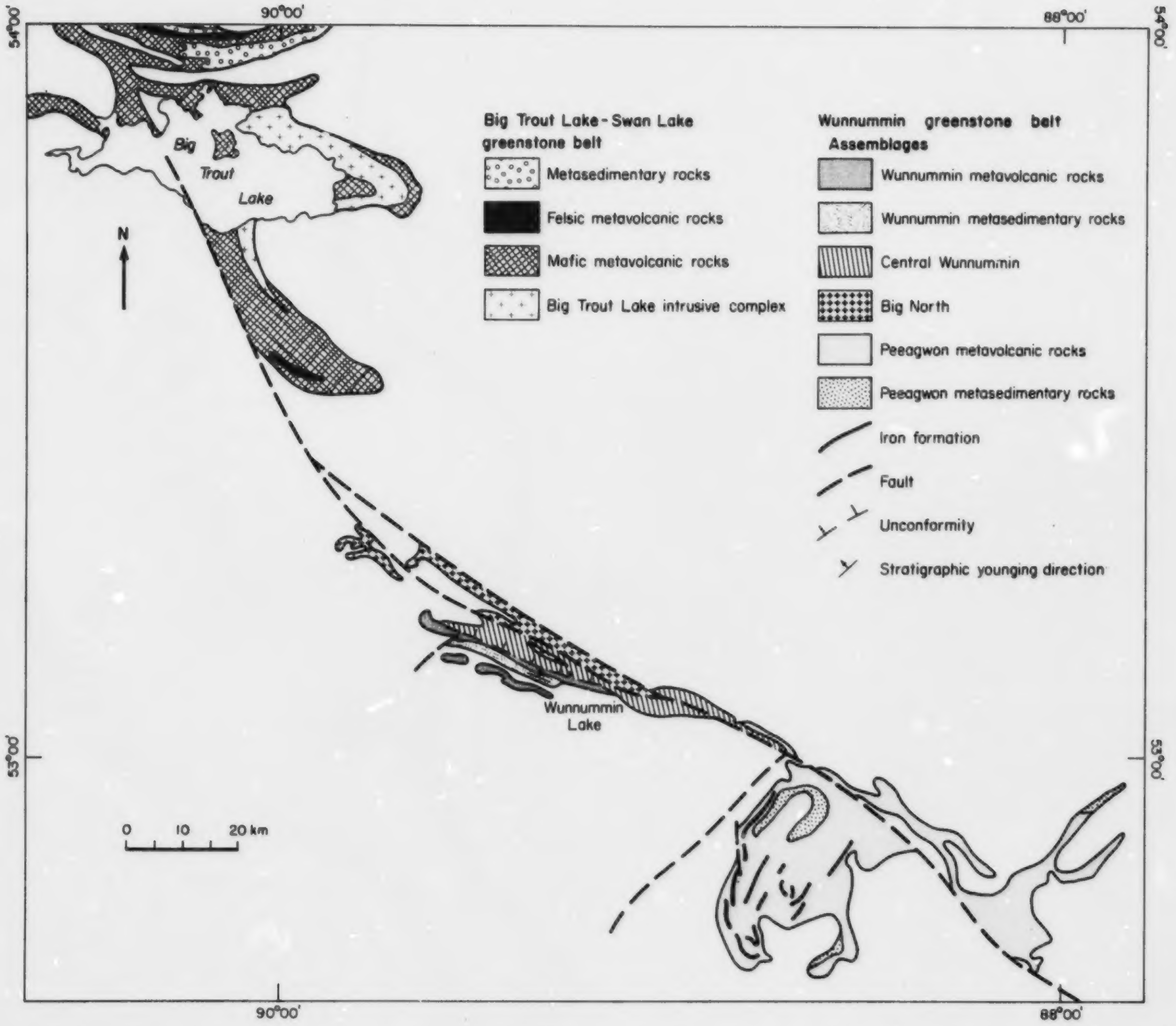

54°00'
90°00'
88°00'
53°00'
Big Trout Lake - Swan Lake greenstone belt
Metasedimentary rocks
Felsic metavolcanic rocks
Mafic metavolcanic rocks
Big Trout Lake intrusive complex
Wunnummin greenstone belt Assemblages
Wunnummin metavolcanic rocks
Wunnummin metasedimentary rocks
Central Wunnummin
Big North
Peeagwon metavolcanic rocks
Peeagwon metasedimentary rocks
Iron formation
Fault
Unconformity
Stratigraphic younging direction
Big Trout Lake
N
Wunnummin Lake
0 10 20 km

Figure 5.8. Sketch of the Big Trout Lake portion of the Big Trout Lake–Swan Lake greenstone belt (*modified from* Hudec 1964; Thurston et al. 1979) and assemblages of the Wunnummin Lake greenstone belt based on the analysis carried out by Thurston et al. (1979) and Prest (1942).

Table 5.5. Assemblages* of the Wunnummin Lake greenstone belt.

Name	Rock Types, Primary Structures (stratigraphy from base to top)	Interpreted Environment	Age (Ma)	References
22) Wunnummin	~ 900 m wide unit of pillowed and massive komatiitic flows with intercalated quartz wacke, quartz arenite and conglomerate; underlain by quartz-rich wacke up to 6.5 km wide; lower contact is intrusive; upper is erosional	platform	2965 detrital	Stott and Janes 1985 Thurston et al. 1987
23) Central Wunnummin	polymictic framework-supported conglomerate with sandstone interbeds; fines upward to sandstone and shale; clasts up to 1.2 m, rounded; clast types = granitic, amphibolite, metasediments (wackes) and rare metavolcanic rocks; disconformably overlies Wunnummin	late unconformable sequence	2710(?)	Prest 1942
24) Big North	pillow and massive mafic flows, minor zones of saussuritization and felsic tuffs; in tectonic contact with Stull Lake–Wunnummin Lake fault zone; uncertain contact relationship with Central Wunnummin	arc volcanics	2.8 Ga(?)	Thurston et al. 1979
25) Peeagwon	pillowed and massive mafic flows, minor units of oxide + sulphide-facies iron formation; interbeds of quartzite and felsic tuff	platform	2.9 Ga(?)	—

The numbering format for assemblages within Table 5.5. corresponds to that used in Figure 5.2.

River Mine at Foster Lake. Felsic metavolcanic rocks varying from lapillistone to tuff breccia occur at Ellard Lake and Gummer Lake. Metasedimentary units include 30 to 300 m wide lenses of boulder conglomerate near the east end of the belt, and near Ellard Lake.

The belt has furnished no stratigraphic younging information. Given the spatial relationship with the Big Trout Lake greenstone belt, the Blackbear River–Ellard Lake greenstone belt may have a similar age. The belt is terminated at the west end by the SLWLFZ, and is intersected by subsidiary shear zones related to the South Kenyon Fault Zone (Osmani and Stott 1988). The South Kenyon Fault Zone may control the location of the lode gold deposit at Foster Lake (Meen 1938).

Miscellaneous Greenstone Belts

Several small areas of greenstone, mainly mafic metavolcanic rocks of medium metamorphic grade, occur throughout the subprovince, including the east end of the Island Lake greenstone belt. In most cases, these areas of greenstone represent remnants commonly correlative with nearby larger greenstone occurrences.

BERENS RIVER SUBPROVINCE GREENSTONE BELTS

The greenstone belts of the Berens River Subprovince form a train of kilometre-scale greenstone remnants and small greenstone belts extending north from the Red Lake and Birch–Uchi greenstone belts of the Uchi Subprovince described by Card and Ciesielski (1986). These relatively small belts exhibit lithostratigraphic associations and structural trends similar to belts of the northern part of the Uchi Subprovince. The larger belts are described below.

McInnes Lake Greenstone Belt

This belt (*see* Figure 5.2) has been mapped by Averill and Ayres (1968) and Cortis et al. (1988). The belt is a medium- to high-grade collage of 3 tectonically bounded assemblages (Figure 5.15a, Table 5.10) with metamorphism increasing to sillimanite grade in proximity to shear zones. The assemblages exhibit complex younging relationships, but the Western McInnes assemblage (Cortis et al. 1988) is the oldest, based on 1) the presence of abundant quartz arenite and quartzose wacke interbeds, similar to the 2.9 billion-year-old greenstones to the north, and 2) the fact that the assemblage is cut by dikes lithologically correlated with distinctive basaltic rocks of the Power assemblage. The Power assemblage is intermediate in age, given the above relationship of Power assemblage mafic dikes cutting Western McInnes assemblage. The Power assemblage is considered to be 2.89 to 2.83 billion years of age, by comparison of ages and rock types with dated sequences at Red Lake (Corfu and Wallace 1986) and Favourable Lake (Corfu and Ayres, in press). The Southern McInnes assemblage is of uncertain age, possibly 2.8 billion years old, based on arc volcanism of similar age to the north (Corfu and Ayres, in press) and south (Corfu and Wallace 1986).

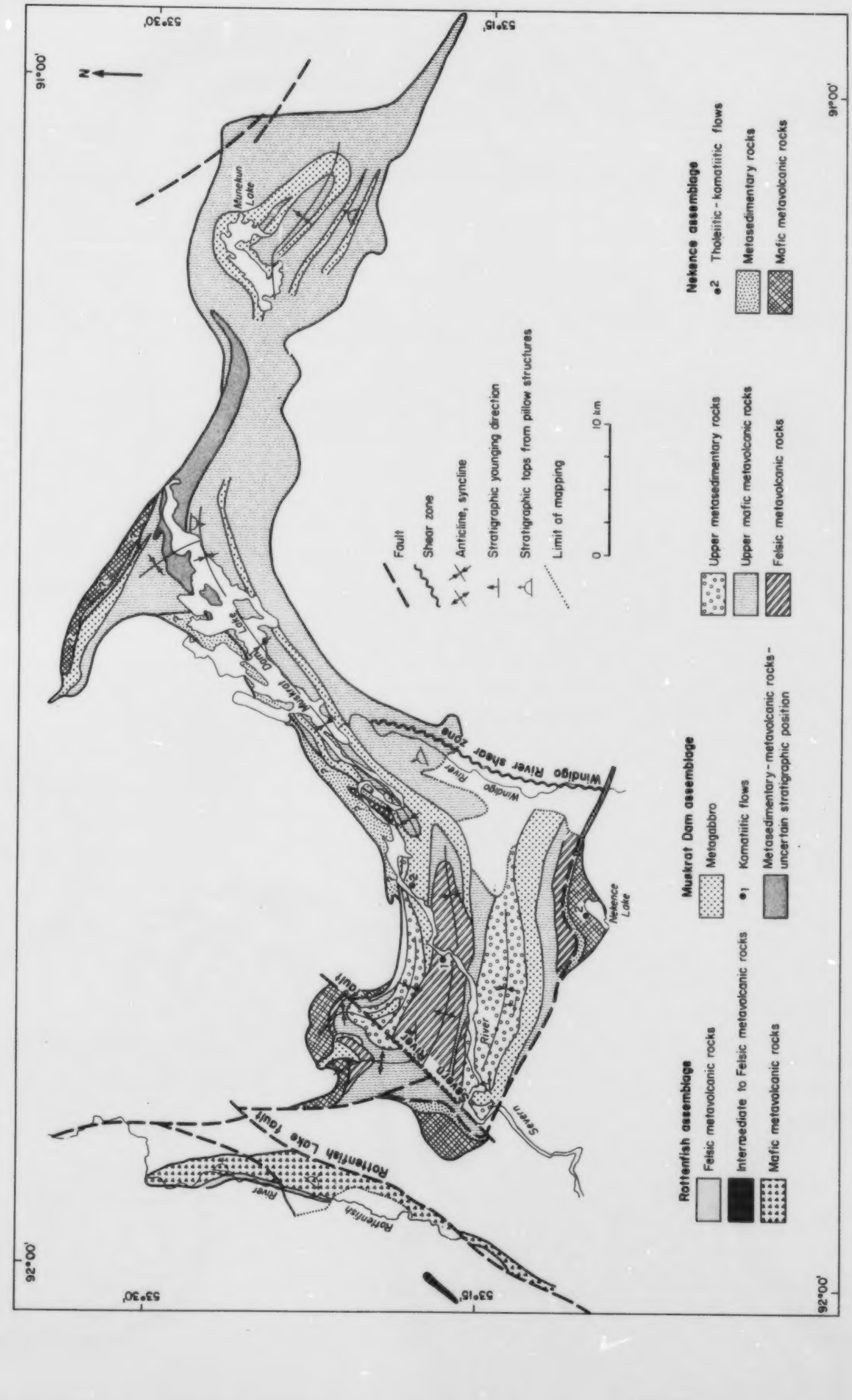

91°00'
53°30'
53°15'
92°00'
N
Minekun Lake
Muskrat Dam Lake
Windigo River shear zone
Windigo River
Nekence Lake
Severn River
Rottenfish Lake fault
Rottenfish River
Fault
Shear zone
Anticline, syncline
Stratigraphic younging direction
Stratigraphic tops from pillow structures
Limit of mapping
0
10 km
Rottenfish assemblage
Felsic metavolcanic rocks
Intermediate to Felsic metavolcanic rocks
Mafic metavolcanic rocks
Muskrat Dam assemblage
Metagabbro
•1 Komatiitic flows
Metasedimentary - metavolcanic rocks - uncertain stratigraphic position
Upper metasedimentary rocks
Upper mafic metavolcanic rocks
Felsic metavolcanic rocks
Nekence assemblage
•2 Tholeiitic - komatiitic flows
Metasedimentary rocks
Mafic metavolcanic rocks

Figure 5.9. Assemblages of the Muskrat Dam Lake greenstone belt based on mapping carried out by Ayres (1969) and Thurston et al. (1987).

Table 5.6. Assemblages* of the Muskrat Dam Lake greenstone belt.

Name	Rock Types, Primary Structures (stratigraphy described from base to top)	Interpreted Environment	Age (Ma)	References
26) Nekence	basal massive and pillowed tholeiitic flows; minimum 200 m thickness of pillowed komatiitic flows near Necklace Lake; overlain by iron formation with metre-scale quartz arenite interbeds, aluminous cummingtonite(?)-bearing clastic metasediments; 1 m wide conglomerate with quartz arenite matrix, felsic metavolcanic and chert clasts; uppermost unit is marble + calc-silicate and a has sheared upper contact	platformal	2958 to 2967 detrital	Ayres 1969 Thurston et al. 1987
27) Muskrat Dam	Fox Bay syncline basal quartz + feldspar-phyric flows (2734 Ma), within a ~2793 Ma inherited component; above are amygdular andesitic flows, pyroclastics capped by a unit of reworked, bedded, graded, well-sorted, andesitic coarse pyroclastics fining up to medium-bedded silts with arenites and conglomerates	arc volcanism	2734	
	Sandhill Crane anticline tuffs with minor ash-flow on Severn River; upper metasediments are reworked andesite tuff to tuff breccia; well-developed bedding, grading, sorting; overlain by graded silts, arenites, conglomerates		2734	
	East of Windigo River shear zone South limb of syncline — upper mafic unit i) mafic pillowed + massive slightly amygdular flows; inter-pillow chert, pyrite ii) overlain by metasedimentary-metavolcanic unit of uncertain position; conglomerates and wacke—heterolithic pebble to cobble clast-supported conglomerate; clasts = felsic metavolcanics, chert, wacke matrix, wacke + silt interbeds			Ayres 1969
	North limb of syncline — basal pillowed basalt unit overlain by felsic metavolcanics (medium- to fine-grained pyroclastic rocks), and mafic flows; cut by a gabbro sill			
	Lower metasedimentary formation wackes and conglomerates lacking quartz arenite clasts			
28) Rottenfish	east margin = Rottenfish Shear Zone; west margin = intrusive contact; basal pillowed and massive mafic flows overlain by felsic flows, tuff and minor coarse pyroclastic rocks	arc volcanism	2.9 Ga(?)	Davis, personal communication, 1990

***The numbering format for assemblages within Table 5.6 corresponds to that in Figure 5.2.**

Table 5.7. Assemblages* of the Sandy Lake greenstone belt.

Name	Rock Types, Primary Structures (stratigraphy described from base to top)	Interpreted Environment	Age (Ma)	References
29) North Sandy	3000 to 5000 m wide; north contact = intrusive, south contact = Sandy Lake shear zone i) pillowed and massive mafic flows (30 to 100 m thick); interflow chert, ferruginous chert iron formation form unit up to 4 km wide ii) a felsic centre with base surge, intra- and extra-caldera facies ignimbrites, cloud surge and air-fall deposits range from pyroclastic breccia to tuff; subaerial deposition iii) oxide-facies iron formation	arc volcanism	2945 crystallization	Satterly 1938 Thurston et al. 1987 Davis, personal communication, 1990
30) Keewaywin	maximum 2000 m wide tectonic contact with North Sandy; i) proximal unit of medium bedded quartzite, conglomeratic sandstones and conglomerates, slump deposits with convolute bedding, sheet sandstones, parallel laminations ii) medial unit of sandstones, muds and silts, abundant scour structures; sand:mud > 1 iii) distal unit of sand:mud < 1; sands, muds, scour structures iv) deep basinal unit of silicate iron formation, thinly laminated graphitic mudstones, diopside carbonates, granular tremolite- and actinolite-rich laminae (tuff?), and dolomitic mudstones	submarine fan complex	2970 to 3048 detrital	Cortis 1988 Davis 1990
31) Sandborn	4000 m wide north contact = Sandy Lake shear zone, south contact = intrusive; i) basal wacke-pelite couplets (A–E turbidites), with 10 to 15 m thick coarse sand base; pelites are magnesium-rich with up to 70% cordierite in a homfels aureole around gabbroic sills ii) pelites overlain by quartzose wacke with argillite conglomerate and lithic wacke interbeds; wacke-argillite couplets up to 2.2 m thick; planar-laminated base 15 cm thick, then up to 15 cm at planar-laminated quartzose wacke capped by 15 cm argillite tops iii) quartzose unit up to 450 m thick succeeded by 6 m of brecciated limestone with no primary structures iv) uppermost unit is 800 m of pillowed komatiitic and tholeiitic flows; unit grades laterally to wacke-pelite unit without volcanics v) turbidites are overlain by 150 to 200 m of felsic quartz and feldspar–phyric ashflow	proximal submarine fan and platform	2.9 Ga (?)	Thurston et al. 1987

Table 5.7. Assemblages* of the Sandy Lake greenstone belt.

Name	Rock Types, Primary Structures (stratigraphy described from base to top)	Interpreted Environment	Age (Ma)	References
32) West Arm	3000 m wide unit of i) tholeiitic and komatiitic subordinate tuff, lapilli tuff (thin-bedded) overlain by quartzose wacke (30 cm to 1 m thick beds with argillite tops) ii) then heterolithic conglomerate; clasts as follows: unit 1 = 30 m thick 10% quartz arenite, 60% fine quartz arenite and chert, 15% amphibolite; unit 2 = 10 to 15% grunerite-magnetite wacke clasts, 45% oxide iron formation, 46% quartz arenite	oceanic volcanism (?)	> 2900	—

* *The numbering format for assemblages within Table 5.7 corresponds to that used in Figure 5.2.*

Table 5.8. Assemblages* of the Pierce–Ponask–Sachigo greenstone belt.

Name	Rock Type, Primary Structures (stratigraphy described from base to top)	Interpreted Environment	Age (Ma)	References
33) Sachigo	– eastern unit of Thurston et al. 1987 i) south–younging basal conglomerates, quartzose wackes and conglomerates are matrix supported, polymictic, 80 to 90% immature quartz in matrix; clasts are as follows: 36% matrix, 5% fuchsitic chert, 3% quartz arenite, 18% mafic metavolcanic, 14% felsic metavolcanic, 16% chert, 8% iron formation; ii) overlain by ultramafic to mafic flows (actinolite-chlorite-talc), plagioclase-phyric mafic vesicular flows, then intermediate pumiceous tuff to lapilli tuff	oceanic volcanism equivalent to Molson Domain, Teacher Group	>2839	Thurston et al. 1987 Corkery and Lenton 1989
34) Pierce	– western unit of Thurston et al. 1987 i) metasedimentary unit = granule to pebble, matrix-supported, oligomict to polymict hornblende-biotite siltstone matrix, rare detrital quartz; 20 to 60% matrix; clasts are 20 to 60% granitoid, 3 to 5% mafic metavolcanic, 5 to 10% angular chert, ii) succeeded to south by pillowed to massive mafic to intermediate flows; felsic thick-bedded plagioclase and quartz-phyric tuff to lapilli tuff ash flows suggesting southward younging; cut by peridotitic sills and a layered ultramafic intrusion at the west end of Ponask Lake	arc volcanism equivalent to Island Lake (?)	>2886	Thurston et al. 1987 Turek et al. 1986

**The numbering format for assemblages within Table 5.8 corresponds to that used in Figure 5.2.*

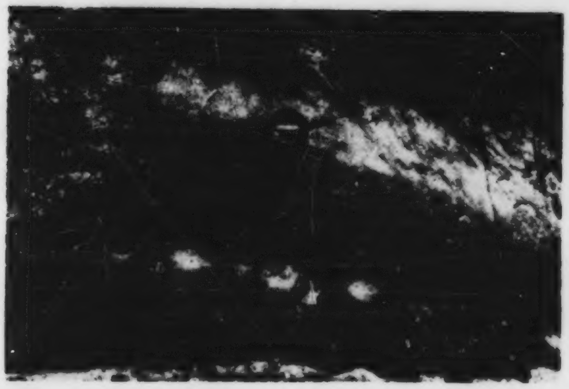


Figure 5.10. Photograph of stromatolites in conglomerate with carbonate matrix at the northwest corner of Sandhill Crane Island on Muskrat Dam Lake.

Hornby Lake Greenstone Belt

This greenstone belt lies in the northern part of the Berens River Subprovince (*see* Figure 5.2). The belt is divided into a Western Hornby assemblage and a Findlay assemblage (Figure 5.15b).

WESTERN HORNBY ASSEMBLAGE

The western part of the Western Hornby assemblage consists of east-younging, pillowed to massive, tholeiitic and komatiitic flows, with subordinate mafic to ultramafic sills and minor wacke, mudstone and sulphide-facies iron formation. Aeromagnetic data indicate a similar geology for the eastern part of the Western Hornby assemblage (*see* Shaded Image of Total Magnetic Field of Ontario map, Map 2584, map case).

FINDLAY ASSEMBLAGE

The Findlay assemblage is sillimanite-bearing wacke interbedded with banded recrystallized chert, mudstone, quartz wacke, quartz arenite, and garnet-biotite wacke. The assemblage is capped by up to 400 m of west-younging dacitic block and ash flows.

The assemblages forming the Hornby Lake greenstone belt young in opposite directions; perhaps representing a tectonically produced "back-to-back" relationship.

Miscellaneous Greenstone Belts

Narrow remnants of more extensive greenstone belts extend north from the Red Lake and Birch Lake areas of the Uchi Subprovince (Stone 1988). The greenstone remnants are made up of grey, mesocratic, amphibolitic gneiss in which relict pillows and fragments are recognizable at Coli Lake and Berens Lake. Inclusions of pillowed mafic metavolcanic rocks, at a scale of 10 to 100 m, occur in gneisses northeast of the Hornby Lake greenstone belt.

PRETECTONIC INTRUSIONS IN GREENSTONE BELTS

Predeformation intrusions in the Sachigo and Berens River Subprovince greenstone belts include small quartz- and feldspar-phyric, intermediate to felsic subvolcanic bodies, and mafic to ultramafic sills and stocks. The felsic to intermediate intrusions form stocks and plutons within virtually every greenstone belt. Prominent examples include the porphyritic microtrondhjemite and microdiorite stocks and sills in the North Spirit Lake greenstone belt (Wood 1980b), and the micro–quartz-diorite to quartz monzonite in the Favourable Lake greenstone belt (Ayres 1974). The geochemistry of these bodies mimics that of the enclosing felsic to intermediate metavolcanic rocks. However, in the North Spirit Lake greenstone belt, the Bijou Point complex (Wood 1977; Smith and Longstaffe 1974) is a stock of shoshonitic biotite granodiorite and quartz diorite with genetically related pyroclastic rocks.

The mafic to ultramafic intrusions within the Sachigo and Berens River Subprovince greenstone belts are synvolcanic, based upon textural and geochronological evidence. In the Favourable Lake area, sills from 17 to 240 m thick intruded non-lithified deposits of the 2926 million-year-old (Corfu and Ayres, in press) caldera-fill sequence of formation I (*see* Table 5.3). These sills represent a tholeiitic basaltic parent magma which has differentiated from pyroxenite upwards into gabbro. The Muskrat Dam Lake greenstone belt contains gabbroic sills up to 300 m thick with a slight upward increase in quartz and a corresponding decrease in mafic minerals (Ayres 1969).

The Big Trout Lake intrusive complex is a large folded sill nearly 5000 m thick whose emplacement predates metamorphism. The complex consists of a lower ultramafic sequence and an upper mafic sequence (*see* Figure 5.8). The lower ultramafic sequence is up to 500 m thick and consists of a peridotite unit with dunite, chromite and chromite-rich layers and an upper homogeneous peridotite. The upper unit includes gabbro, anorthosite and anorthositic gabbro grading up into quartz diorite. Geochemical relationships (Borthwick and Naldrett 1984) indicate that the sill was the product of at least 2 batches of tholeiitic magma.

SACHIGO AND BERENS RIVER SUBPROVINCE GRANITIC ROCKS

Granitic rocks in the Sachigo and Berens River subprovinces have been mapped only on a reconnaissance basis, with greater detail available only in the vicinity of North Caribou Lake (Breaks et al., in prep.), Favourable Lake (Ayres 1974; Corfu et al. 1985), and the northern and southern margins of the Berens River Subprovince (Stone 1988, 1989). Field relationships have shown the following sequence of intrusion of granitic rocks, from oldest to youngest, to hold generally across the region, using the nomenclature of the Bedrock Geology of Ontario maps (*see* Maps 2542 to 2543, map case):

1. the gneissic tonalite suite
2. the foliated tonalite suite

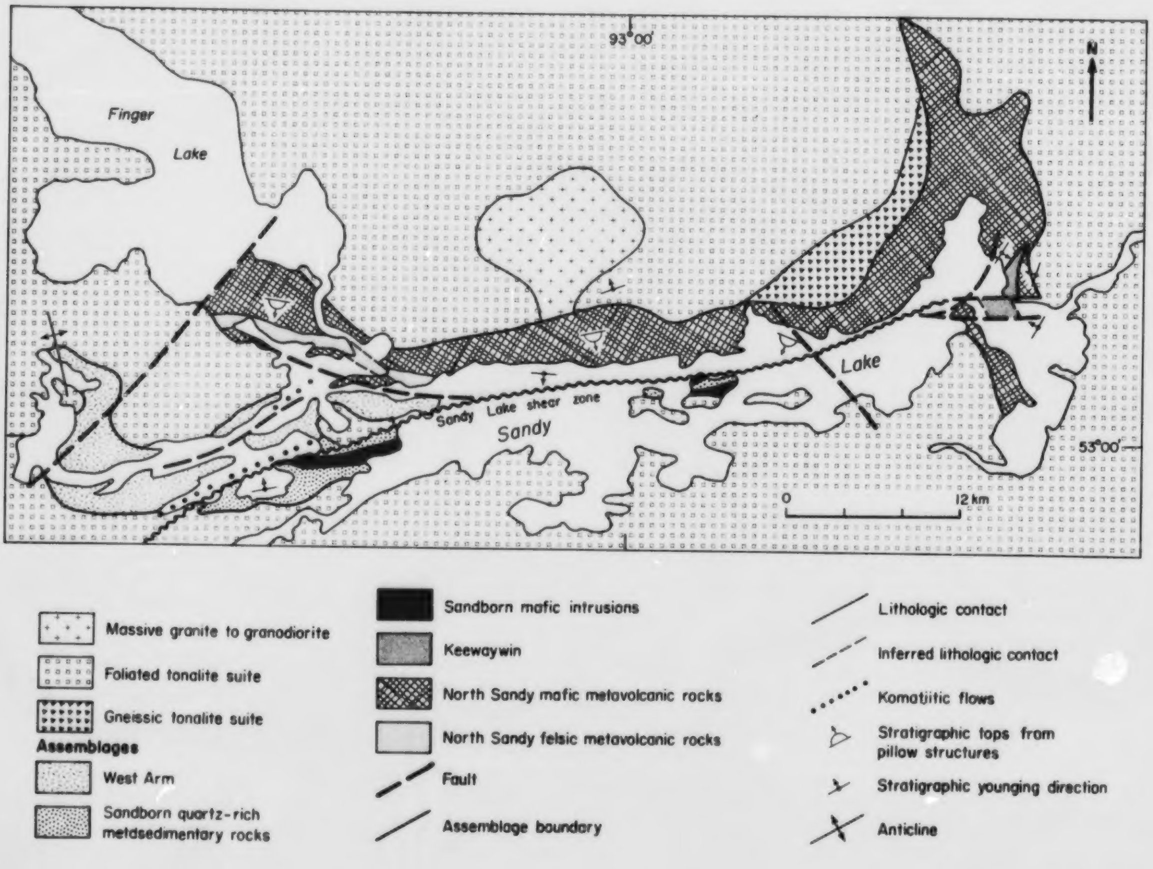
3. the massive granodiorite to granite suite
4. the muscovite-bearing granite suite
5. the diorite-monzonite-granodiorite suite

Examples of suites 1, 2 and 3 are commonly spatially associated (*see* Bedrock Geology of Ontario, Maps 2541 to 2543, map case). Data from the northwestern Superior Province indicate that there may be more than one age of muscovite-bearing granitic rocks. Representative occurrences of each of the major units from both the Berens River and the Sachigo subprovinces will be described in the following section.

Gneissic Tonalite Suite

With the possible exception of the early tonalitic basement complex of Hillary and Ayres (1980) and correlative units, this suite is the oldest plutonic unit in the Sachigo and Berens River subprovinces. We subdivide the unit into 1) mafic tonalite to granodiorite gneiss which generally contains more than 20% hornblende and biotite as well as centimetre- to metre-scale darker enclaves; and 2) tonalite to granodiorite gneiss which is typically more leucocratic than 1.

The gneisses form the bulk of map unit 11 (*see* Bedrock Geology of Ontario, Maps 2451 to 2543, map case) and have contacts that are generally tectonic where adjacent to greenstone belts (Breaks et al., in prep.; Stott and Janes 1985; Ayres 1969; Wilson 1987), and gradational where adjacent to felsic plutonic rocks. Crosscutting relations show that these gneisses are older than granitoids of the sodic and potassic suites of units 12, 14, and 15 (*see* Bedrock Geology of Ontario, Maps 2541 to 2543, map case). The gneissic rocks are generally grey, mesocratic to leucocratic, and compositionally and texturally heterogeneous. Typical exposures are cut by several generations of granitic dikes and may contain kilometre- to metre-scale mafic inclusions. Some



93°00'
N
Finger
Lake
Lake
Sandy Lake shear zone
Sandy
53°00'
0
12 km
Massive granite to granodiorite
Foliated tonalite suite
Gneissic tonalite suite
Assemblages
West Arm
Sandborn quartz-rich metasedimentary rocks
Sandborn mafic intrusions
Keewaywin
North Sandy mafic metavolcanic rocks
North Sandy felsic metavolcanic rocks
Fault
Assemblage boundary
Lithologic contact
Inferred lithologic contact
Komatiitic flows
Stratigraphic tops from pillow structures
Stratigraphic younging direction
Anticline

Figure 5.11. Assemblages of the Sandy Lake greenstone belt based on mapping carried out by Satterly (1939), Thurston et al. (1987) and Cortis (1988).

inclusions, such as those north of Akow Lake in the North Caribou Lake area, preserve textural evidence of an origin as supracrustal units; others have a texture and distribution suggestive of an origin as tectonized mafic dikes. Tonalite to granodiorite gneisses contain less than 20% mafic minerals with biotite dominant; mafic varieties of gneiss are hornblende-bearing with a colour index of 20 to 50. Variable proportions of quartz and feldspar cause this suite to span the range from tonalite to granite. Thin sections (Stone 1990) show blocky elongated and recrystallized quartz and feldspar grains with accessory epidote, chlorite, zircon, apatite and opaque minerals.

The gneissic tonalite suite forms large complex batholiths in Sachigo Subprovince. These batholiths are situated on both sides of the North Caribou greenstone belt, south of the Big Trout Lake–Swan Lake greenstone belt, and east and west of the Sandy Lake greenstone belt. Several sill-like remnants south of Favourable Lake are also composed of gneissic tonalite.

The gneisses have a west- to northwest-striking differentiated layering which diverges around younger oval batholiths and is rotated into parallelism with major shear zones such as the Stull Lake–Wunnummin Lake fault zone (*see* Figure 5.2). Uranium-lead zircon geochronology yields an age of crystallization of 2950 Ma for a tonalite enclave at Favourable Lake (Corfu and Ayres 1984). This technique has yielded metamorphic ages of 2900 Ma on 2 leucocratic gneisses near the south margin of the Berens River Subprovince (Krogh et al. 1974). Given the younger age of most metavolcanic units relative to the gneissic tonalite suite, and the tectonic contacts of the gneisses with greenstone belts, the gneisses may predate volcanism, although no basement-cover relationship has been preserved (Hillary and Ayres 1980).

Foliated Tonalite Suite

Rocks of the foliated tonalite suite form irregular batholiths and sheet-like plutons and include mafic tonalite to granodiorite (Stone 1989), hornblende tonalite to granodiorite, and biotite tonalite to granodiorite. The mafic tonalite to granodiorite is a coarse, white to grey rock with less than 20% clotted mafic minerals and locally intense development of a protomylonitic fabric; these rocks occur as small enclaves in felsic plutonic areas and as regional-scale sheets flanking the western part of the Favourable Lake greenstone belt (Figure 5.16). The protomylonitic fabric tends to occur towards the margins of many of the occurrences of these rocks and may thus represent an artifact of the intrusion process.

The best known mafic tonalite unit occurs within the North Trout Lake batholith described by Hillary and Ayres (1980) (Figure 5.17a). The unit is a small 2950 million-year-old (Corfu et al. 1985) enclave of gneissic tonalite surrounded by younger (2732 to 2711 million-year-old) foliated tonalite and porphyritic granodiorite. Biotite tonalite within the enclave is commonly foliated, but in part gneissic, with up to 10% sill-like melanocratic layers up to 1 m thick and 5% leucocratic metamorphic segregations less

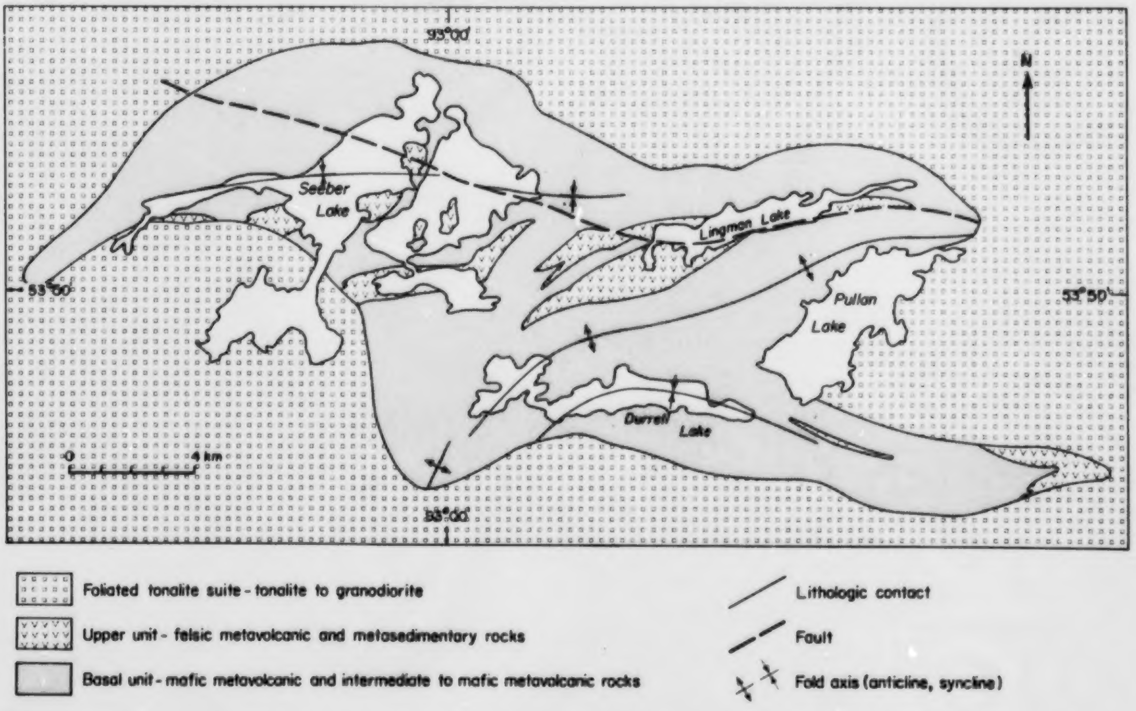

93°00'
N
Seeber Lake
Lingman Lake
53°50'
53°50'
Pullan Lake
Durrell Lake
0
4 km
93°00'
Foliated tonalite suite - tonalite to granodiorite
Upper unit - felsic metavolcanic and metasedimentary rocks
Basal unit - mafic metavolcanic and intermediate to mafic metavolcanic rocks
Lithologic contact
Fault
Fold axis (anticline, syncline)

Figure 5.12. Sketch map of the geology of the Lingman Lake greenstone belt *modified from* Wilson (1987).

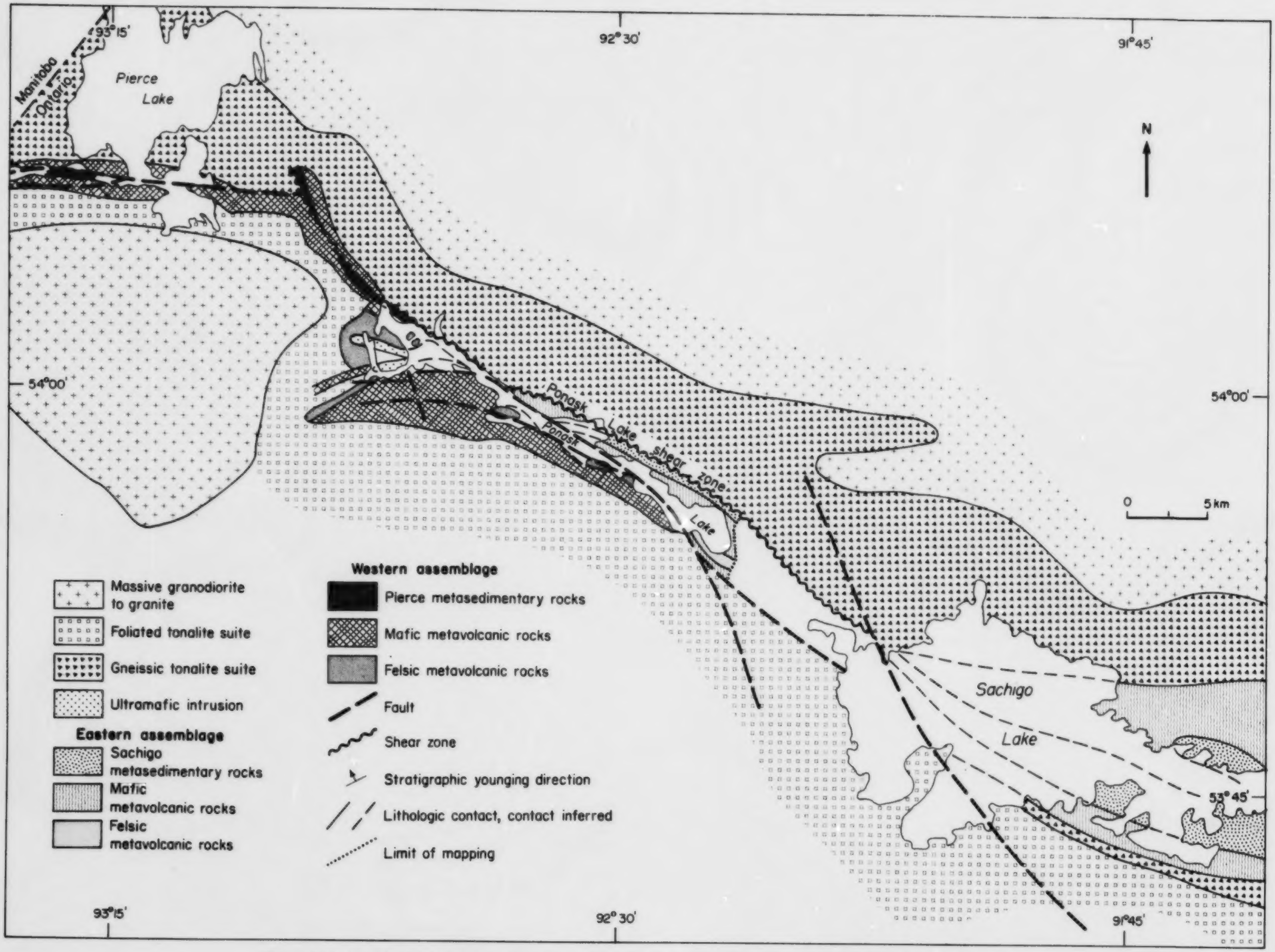

93°15'
92° 30'
91°45'
54°00'
53°45'
Manitobo
Ontario
Pierce
Lake
Ponask
Lake
shear
zone
Ponask
Lake
Sachigo
Lake
N
0
5 km
Massive granodiorite to granite
Foliated tonalite suite
Gneissic tonalite suite
Ultramafic intrusion
Eastern assemblage
Sachigo metasedimentary rocks
Mafic metavolcanic rocks
Felsic metavolcanic rocks
Western assemblage
Pierce metasedimentary rocks
Mafic metavolcanic rocks
Felsic metavolcanic rocks
Fault
Shear zone
Stratigraphic younging direction
Lithologic contact, contact inferred
Limit of mapping

Figure 5.13. Assemblages of the Pierce–Ponask–Sachigo greenstone belt based on mapping done by Thurston et al. (1987).

than 5 cm thick. The tonalite contains amphibolite xenoliths, tonalite dikes, and locally well-preserved igneous plagioclase textures. In general though, the unit is extensively recrystallized.

Foliation in the tonalite is defined by oriented lenticles of quartz, plagioclase, biotite and hornblende; where the aggregates are thickest and most extensively developed, they define a gneissic foliation. The leucocratic layers have the same plagioclase:quartz ratio as the host rock, but are less strained and recrystallized. The melanocratic layers are made up of slightly more mafic tonalite. The geochemistry of the mafic tonalite is presented in Table 5.11 (Hillary and Ayres 1980). Rare earth element geochemistry is portrayed in Figure 5.17b. The rare earth data show the tonalites to be similar to the high Al_2O_3 trondhjemite class described by Arth and Barker (1976).

Hornblende tonalite to granodiorite underlies large parts of the Berens River and Sachigo subprovinces and is partly assimilated by younger granitic units. This rock type is white to grey, and medium to coarse grained, with blocky amphibole and, locally, potassium feldspar megacrysts. The unit has a colour index of about 20, with subequal amphibole and biotite crystals; compositional variation is shown on Figures 5.18a and 5.18b.

Biotite tonalite to granodiorite forms distinct intrusions (Stone 1989) and phases within hornblende-biotite tonalite batholiths such as the North Trout Lake batholith (Ayres 1974). This rock type is white to grey and medium grained, with 5 to 15% biotite; variations in leucocratic minerals are shown in Figure 5.18. Although some intrusions have a single phase of biotite tonalite, others show many phases. Ayres (1974) mapped 20 phases in the North Trout Lake batholith based on grain size, texture, composition, size, shape and abundance of phenocrysts, abundance of quartz, abundance of mafic minerals, hornblende:biotite ratio, structure and colour (Figure 5.19).

The intrusions are moderately well foliated and form distinct stocks and batholithic complexes. The intrusions are commonly crescentic plutons (Stott 1986) which intrude the interface between gneissic tonalites of unit 11 (*see* Bedrock Geology of Ontario, Maps 2541 to 2543, map case) and greenstone belts; for example, along the southwest margin of the North Caribou greenstone belt. Batholithic complexes such as the North Trout Lake batholith include sodic and potassic plutonic rocks and gneisses (Ayres 1969). Many of these plutons have imposed metamorphic aureoles upon neighbouring greenstone belt units (e.g., Breaks et al., in prep.; Ayres 1969, 1974, 1978; Cortis et al. 1989). The absolute ages of these rocks may be quite variable: the North Caribou batholith (Breaks et al., in prep.) is 2871 million years old (Davis 1987 in deKemp 1987); and the range in ages of 2732 to 2711 Ma is obtained for the North Trout Lake batholith (Ayres 1974; Corfu et al. 1985), and the Little Vermilion Lake batholith (Corfu and Andrews 1987). The

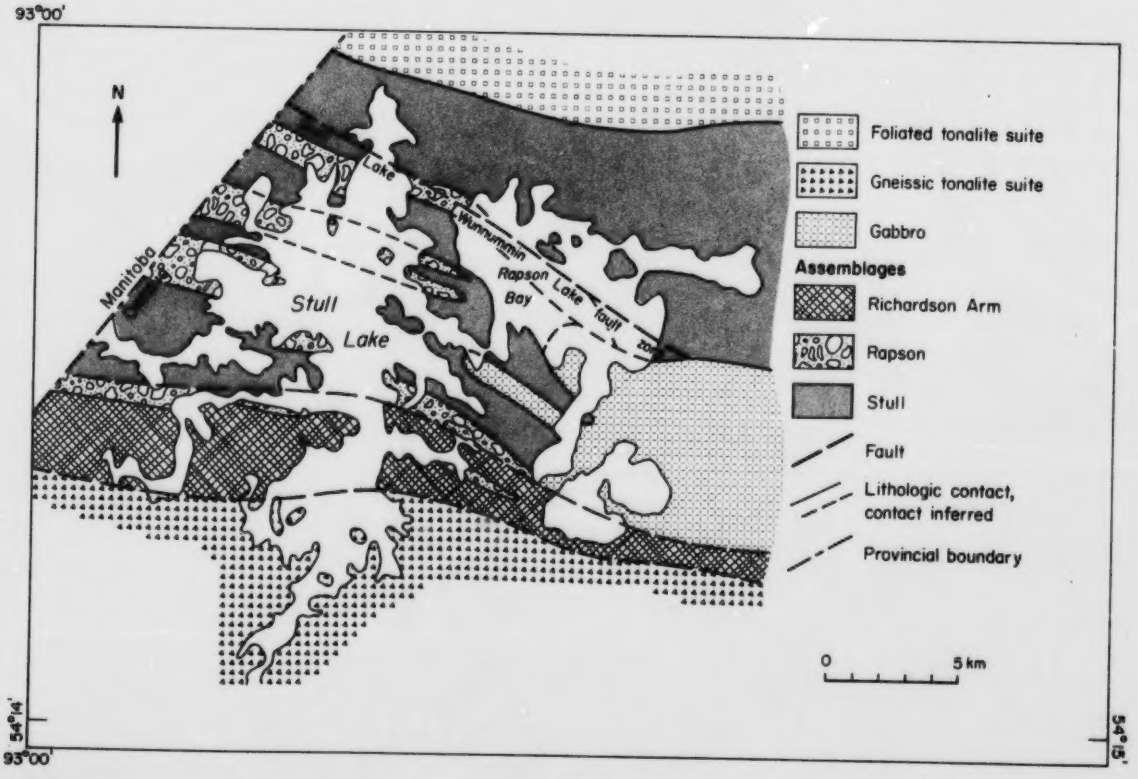


Figure 5.14. Assemblages of the Stull Lake portion of the Big Trout Lake–Swan Lake greenstone belt based upon mapping carried out by Satterly (1938) and supplemented by Thurston et al. (1987).

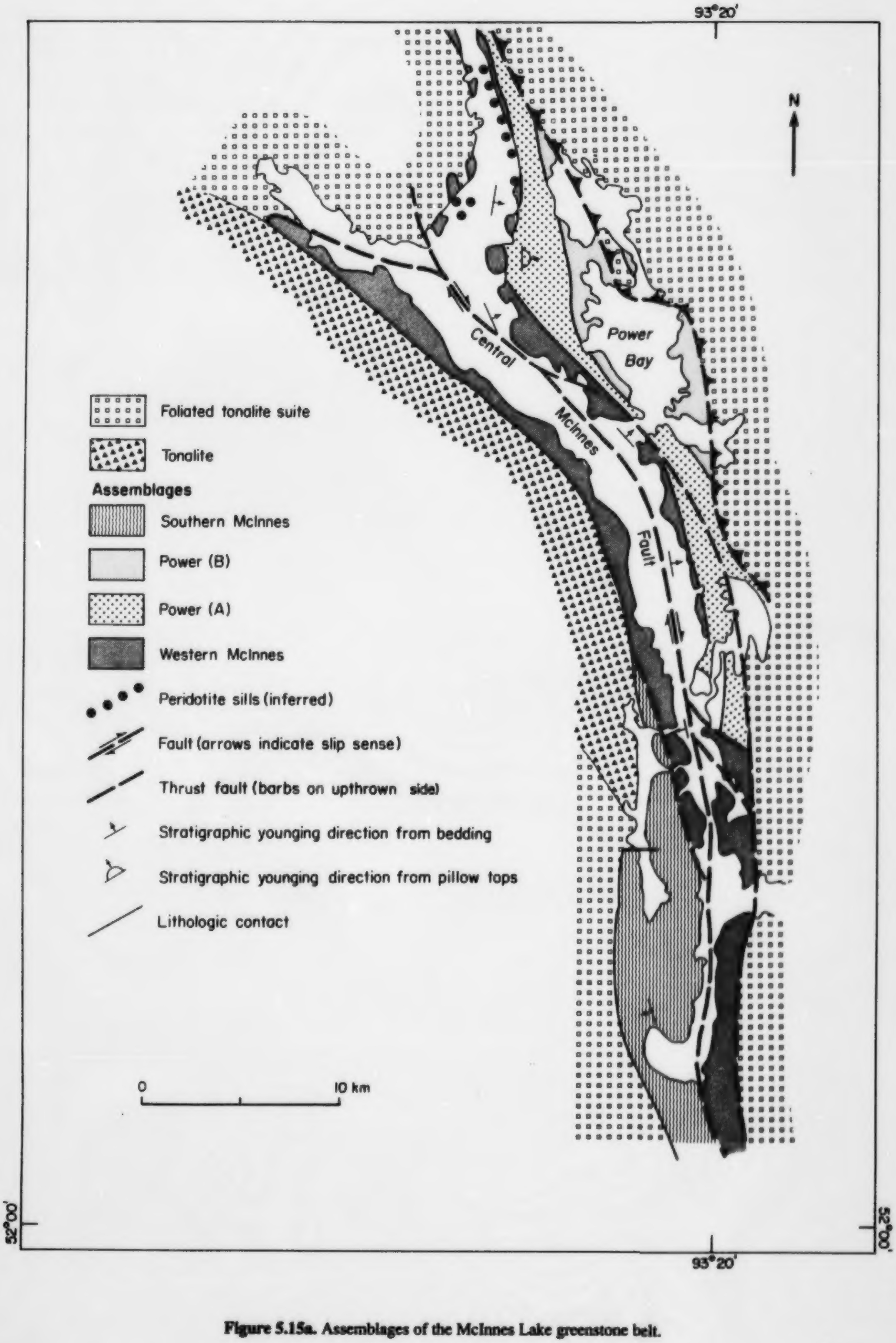

93°20'
N
Power Bay
Central McInnes Fault
Foliated tonalite suite
Tonalite
Assemblages
Southern McInnes
Power (B)
Power (A)
Western McInnes
Peridotite sills (inferred)
Fault (arrows indicate slip sense)
Thrust fault (barbs on upthrown side)
Stratigraphic younging direction from bedding
Stratigraphic younging direction from pillow tops
Lithologic contact
0
10 km
52°00'
52°00'
93°20'

Figure 5.15a. Assemblages of the McInnes Lake greenstone belt.

quartz diorite at MacDowell Lake is 2744 million years old (Corfu and Wood 1986).

Massive Granodiorite to Granite Suite

The most abundant rocks of the massive granodiorite to granite suite are, in order of decreasing relative age, potassic feldspar megacrystic granodiorite to granite, and biotite granodiorite to granite. The megacrystic rocks are grey to pink, typically containing 1 to 2 cm potassic feldspar megacrysts and 10% to 20% biotite and amphibole; variation in the major minerals is shown in Figure 5.18a and 5.18b. The megacrystic rocks vary in composition from granodiorite to granite and are in gradational contact with hornblende tonalite to granodiorite (e.g., the margins of the Little Vermilion Lake batholith) (Stone 1990). Biotite and amphibole occur with opaque oxides in mafic aggregates cored by irregular, inclusion-rich amphibole grains, possibly representing relict clinopyroxene. The opaque mineral is likely magnetite, given the consistently high aeromagnetic expression of this unit throughout the Berens River and Sachigo subprovinces (*see* Shaded Image of Total Magnetic Field of Ontario, Map 2584, map case). Radiometrically dated examples of this unit are 2711 million years old at Setting Net Lake (Corfu and Ayres 1984) and North Trout Lake (Corfu et al. 1985).

Biotite granodiorite to granite is the most extensive rock type of the massive granodiorite to granite suite in the Berens River Subprovince (Stone 1990). The unit is pale pink, weakly foliated and contains irregular pegmatite

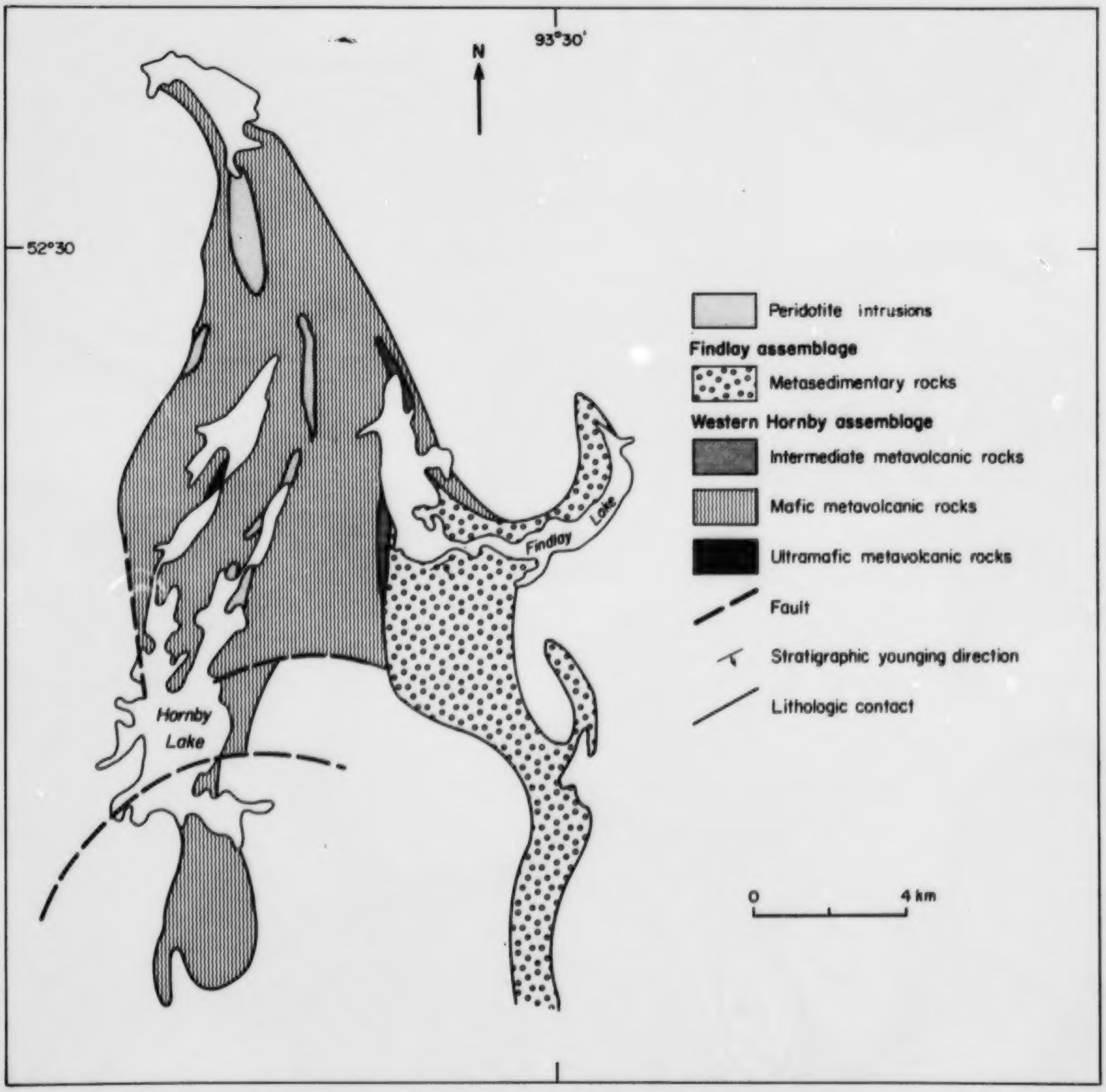

93°30'
N
52°30
Peridotite intrusions
Findlay assemblage
Metasedimentary rocks
Western Hornby assemblage
Intermediate metavolcanic rocks
Mafic metavolcanic rocks
Ultramafic metavolcanic rocks
Fault
Stratigraphic younging direction
Lithologic contact
Findlay Lake
Hornby Lake
0
4 km

Figure 5.15b. Assemblages of the Hornby Lake greenstone belt based on mapping carried out by Cortis et al. (1988).

Table 5.9. Assemblages* of the Stull Lake portion of the Big Trout Lake–Swan Lake greenstone belt.

Name	Rock Type, Primary Structures (stratigraphy described from base to top)	Interpreted Environment	Age (Ga)	References
35) Stull	pillowed and massive mafic flows, minor mafic fragmental units of uncertain origin, intermediate tuff, lapilli tuff, tuff breccia and felsic tuff; most coarse units are heterolithic, matrix-supported units with well-developed grading	arc volcanism correlated with Hayes River Group	2.8	M E M 1987 Satterly 1938
36) Rapson	framework-supported, heterolithic, pebble to boulder conglomerate; typical clast content is 75% granitic, 20% arenaceous metasediments, 5% porphyritic basalt; abundant arkose interbeds	"Timiskaming" type late unconformable sequence correlated with Oxford Lake Group	2.7	
37) Richardson Arm	hornblende porphyritic massive and pillowed mafic flows	correlated with Munro Lake Terrane and Pipestone Group at Cross Lake	2.76(?)	Corkery and Lenton 1989

**The numbering format for assemblages within Table 5.9 corresponds to that used in Figure 5.2.*

Table 5.10. Assemblages* of the McInnes Lake greenstone belt.

Name	Rock Type, Primary Structures (stratigraphy described from base to top)	Inferred Environment	Age (Ga)	References
38) Western McInnes	– Cycle I of Cortis et al. (1988) i) east-younging sequence with 20% basal pillowed(?) basalts overprinted with hornblende porphyroblasts, overlain by quartz- and feldspar-phyric tuffs with minor pyroclastic breccia ii) upper unit includes sandstone-mudstone couplets with Bouma BDE and BE divisions totalling 0.5 to 1 m thick and 1 to 2 m thick polymictic, matrix-supported conglomerates with 99% metavolcanic clasts iii) quartz arenite and quartz wacke beds occur in the upper part of the assemblage	arc volcanism	2.9	Cortis et al. 1988
39) Power	– Cycle II of Cortis et al. (1988) i) east-younging assemblage (II) is 30 m thick basal oxide iron formation then 800 to 900 m of pillowed and massive mafic and ultramafic flows; flows to 40 m thick, up to 30% 1 to 2 cm varioles ii) upper unit (IIa) is up to 1000 m of bedded graded pyroclastic units fining upward from pyroclastic breccia to tuff; bed thickness 0.1 to 10 m; minor turbiditic wacke interbeds	shallow water primitive arc volcanism	2.8(?)	Cortis et al. 1988
40) Southern McInnes	– Cycle III of Cortis et al. (1988) west-younging sequence of quartz- and feldspar-phyric, thinly to very thickly bedded tuffs and crystal tuffs, minor mafic flows, plagioclase-phyric ash flows overlain by 200 to 300 m of mafic flows and tuffs	arc volcanism	?	Cortis et al. 1988

** The numbering format of the assemblages within Table 5.10 corresponds to that used in Figure 5.2.*

zones. Biotite and subordinate amphibole rarely exceed 10% of the rock; variation of major minerals is displayed in Figure 5.18. Primary igneous textures are preserved and only minor sericitic alteration is present. A comparable unit at Horseshoe Lake in Manitoba is dated at 2715 Ma (Ermanovics and Wanless 1983), and granite in the Bear Head batholith is dated at 2696 Ma (Corfu and Ayres 1984).

The massive granodiorite to granite suite forms large lobate intrusions and crescentic bodies near granite-greenstone belt contacts in the southern Berens River Subprovince and throughout the Sachigo Subprovince. In the northern part of the Berens River Subprovince, this suite forms west- to northwest-striking elongate sheet-like intrusions.

Muscovite-Bearing Granite Suite

This rock suite comprises coarse-grained to pegmatitic white granodiorite to granite with abundant, partly assimilated inclusions of metasediment. Muscovite, with or without biotite, and accessory garnet, cordierite, apatite and tourmaline are typically present. These rocks include two-mica granite and leucogranite. At some localities, units of muscovite granite have aureoles of pegmatite dikes which are fractionated, based upon chemical and mineralogical criteria. Occurrences of rare-metal minerals in the pegmatite dikes and sheets of two-mica granite are spatially associated (Breaks and Osmani 1989) with major shear zones in northern Sachigo Subprovince and along the Sachigo–Berens River Subprovince boundary (Stone 1989) (*see* Figure 5.16). The intrusions generally lie within the shear zones and are parallel to the margins of the enclosing, or nearby, shear zones. Modal variation for some examples is shown in Figure 5.18. Additional occurrences of two-mica granites and associated pegmatites are found along shear zones within the North Caribou greenstone belt (Breaks et al. 1986) and along many of the major shear zones within the Berens River and Sachigo subprovinces (Breaks and Osmani 1989) (Figure 5.20).

In the southern Superior Province, two-mica granites form stocks and batholiths largely within, and adjacent to, metasedimentary subprovinces (cf. Williams, this volume; Breaks, this volume). These plutons, derived from melting metasedimentary rocks (Ayres and Cerny 1982), are syntectonic to posttectonic, depending upon crustal level (*see* Williams, this volume). In contrast, two-mica granites in the northwestern Superior Province form small stocks associated with regional-scale shear zones (Breaks and Osmani 1989). These northwestern Superior Province two-mica granites, based on sparse data in Manitoba (Mezger et al. 1990), are about 2653 million years old; younger than most other granitic units in the northwestern Superior Province. Therefore, northwestern Superior Province two-mica granites may represent the melting of a

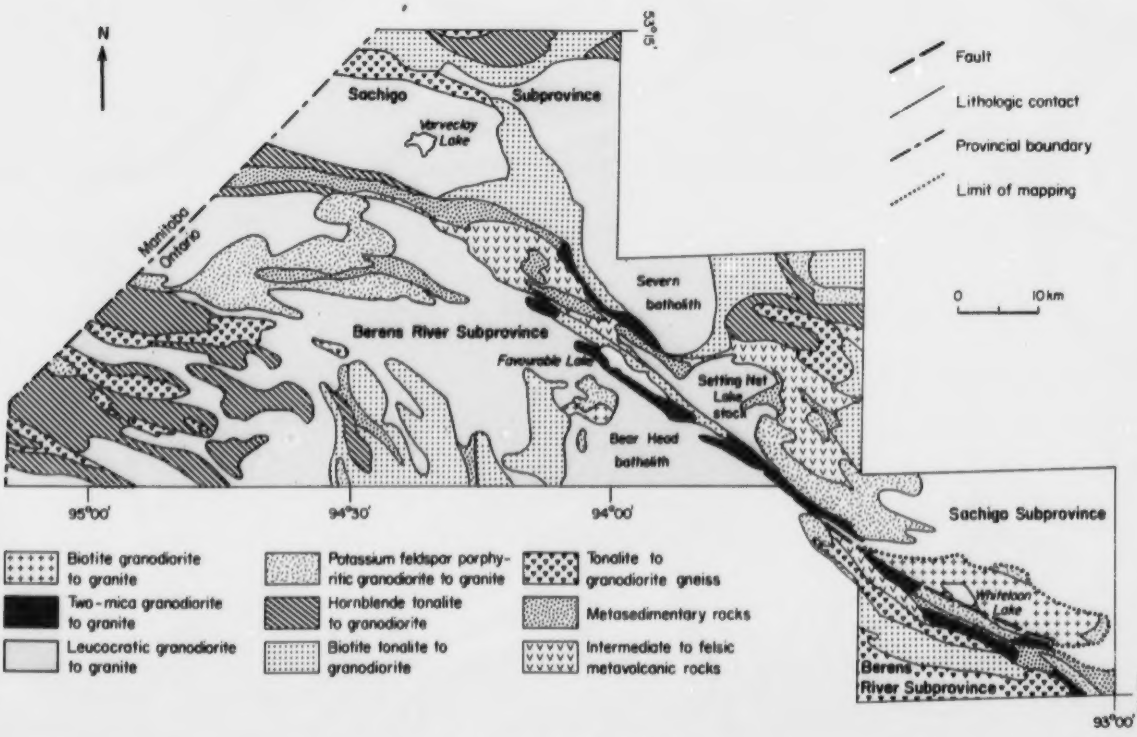

N
53°15'
Sachigo
Subprovince
Varveclay Lake
Manitoba
Ontario
Severn batholith
Berens River Subprovince
Favourable Lake
Setting Net Lake stock
Bear Head batholith
95°00'
94°30'
94°00'
Sachigo Subprovince
Whiteloon Lake
Berens River Subprovince
52°30'
93°00'
Fault
Lithologic contact
Provincial boundary
Limit of mapping
0
10km
Biotite granodiorite to granite
Two-mica granodiorite to granite
Leucocratic granodiorite to granite
Potassium feldspar porphyritic granodiorite to granite
Hornblende tonalite to granodiorite
Biotite tonalite to granodiorite
Tonalite to granodiorite gneiss
Metasedimentary rocks
Intermediate to felsic metavolcanic rocks

Figure 5.16. Map of granitoids of northern Berens River Subprovince (*from* Stone 1990). Note the spatial association of muscovite-bearing granites with the subprovince boundary described by Card and Ciesielski (1986). Card and Ciesielski (1986) used the Favourable Lake greenstone belt as the boundary between the Berens River and Sachigo subprovinces west of the Setting Net Lake stock.

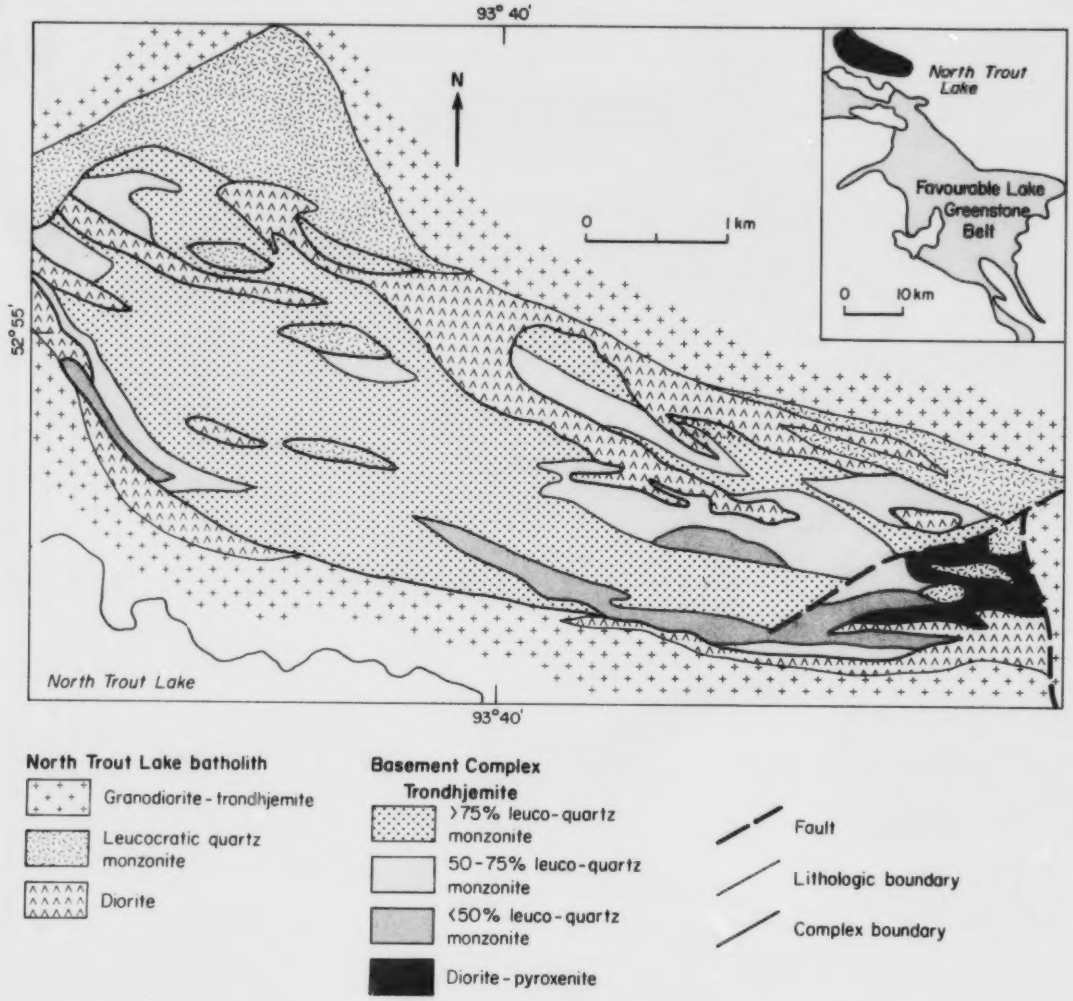

93° 40'
N
0
1 km
North Trout Lake
Favourable Lake Greenstone Belt
0
10 km
52°55'
North Trout Lake
93°40'
North Trout Lake batholith
Granodiorite - trondhjemite
Leucocratic quartz monzonite
Diorite
Basement Complex
Trondhjemite
>75% leuco-quartz monzonite
50-75% leuco-quartz monzonite
<50% leuco-quartz monzonite
Diorite - pyroxenite
Fault
Lithologic boundary
Complex boundary

Figure 5.17a. Map of the Favourable Lake tonalite enclave.

metasedimentary precursor during late block-to-block interaction.

Diorite-Monzonite-Granodiorite Suite

Intrusions of this suite appear to be relatively young compared to nearby felsic plutonic rocks. In southern Berens River Subprovince, a small hornblende granodiorite unit north of Kirkness Lake (Stone 1989) cuts massive granodiorite to granite. The Frame Lake pluton (*see* Figure 5.2) and the Bear Head batholith are dated at 2696 Ma and 2697 Ma respectively (Corfu and Ayres 1984), with no evidence of an inherited older component in the zircon populations. Recent mapping and sampling have increased the number and compositional range of examples of this suite (Stern et al. 1989; Sutcliffe et al. 1990) which may form up to 20% of Archean granitic rocks.

Rock compositions vary from quartz diorite to quartz monzonite with 20% to 30% hornblende and subordinate biotite. The suite characteristically has much more hornblende than biotite, contains centimetre- to metre-scale mafic inclusions, and displays a spatial association with mafic to ultramafic plutons, dikes, layers and enclaves. In rare cases, pyroxene is the major mafic mineral in units of this suite (Stone 1989). The suite grades into mafic members of gabbroic composition, and gradations from the predominant granodiorite and granite to quartz syenite, quartz monzonite and quartz monzodiorite have been reported (Stone 1989; Thurston et al. 1979). In the Berens River Subprovince, the Frame Lake pluton is concentrically zoned, with a diorite rim and a hornblende granodiorite core (Corfu and Ayres 1984). The plutons of this suite, because of their elevated content of mafic inclusions and hornblende, commonly form positive aeromagnetic anomalies (*see* Shaded Image of Total Magnetic Field of Ontario map, Map 2584, map case).

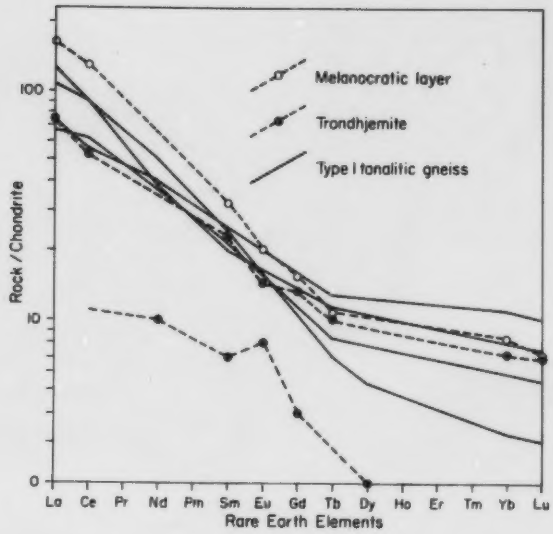


Figure 5.17b. Rare earth element geochemistry of the Favourable Lake tonalite, *modified from* Hillary and Ayres (1980).

Mafic inclusions and dikes are prominent within rocks of this suite east of Wunnummin Lake (Thurston et al. 1979), at Frame Lake (Corfu and Ayres 1984), and at Muskrat Dam Lake (Ayres 1969). The mafic mineralogy, the inclusion-rich nature of the unit, and the lack of inheritance in the geochronologic data suggest rocks of this unit may have originated in a manner similar to monzodiorite (sanukitoid) plutons in the southern Superior Province which are thought to be mantle derived (Stern et al. 1989; Sutcliffe et al. 1990).

WINISK SUBPROVINCE

This subprovince consists of all Archean rocks east of the Winisk River fault. The region is underlain by clinopyroxene-bearing quartz monzonite (Thurston et al. 1979, Appendix 1), granodioritic gneisses and undeformed granitic rocks including granodiorite and minor granite, quartz diorite, and quartz monzonite (Bostock 1971). The subprovince was correlated with the Pikwitonei Subprovince of Manitoba (Ayres et al. 1969) based upon the similar curvilinear aeromagnetic pattern with alternating highs and lows in rocks east of the Winisk River fault. The subprovince is known through exploration diamond drilling in the vicinity of the fault (*see* Figure 5.2) and a small exposed area in the Sutton Inlier (Bostock 1971).

METAMORPHISM

Details of metamorphic assemblages are available for the following areas: Muskrat Dam Lake (Ayres 1969), Favourable Lake (Ayres 1974), North Spirit Lake (Wood 1977, 1980b) and North Caribou Lake (Breaks et al., in prep.). Pretectonic units have been subjected to at least 2 metamorphic events. The greenstone belts display low to upper greenschist-facies interiors and outer amphibolite-facies zones rimmed by thin hornblende hornfels-facies zones where they are adjacent to the younger batholiths (Ayres 1969, 1974) (Figure 5.21). In general, a greenschist-facies interior zone is present in all greenstone belts wider than 5 km within the Sachigo Subprovince (Ayres 1978).

The North Caribou greenstone belt contains mineral assemblages which represent low-pressure intermediate-type metamorphism (Miyashiro 1961) under conditions of 400° to 500°C and pressures of 300 MPa or less (Breaks et al., in prep.). The southeastern part of the greenstone belt exhibits medium-grade metamorphism compared to the low-grade assemblages of the northwestern part of the belt.

The greenstone belts of the Sachigo and Berens River subprovinces have been subjected to multiple metamorphic stages. This has been interpreted on the basis of an early planar fabric preserved as inclusion trails in garnet, and was followed by a later regional metamorphic stage with batholith-related contact aureoles superimposed on the second regional stage (Ayres 1978; Breaks et al., in prep.). Relationships between deformation and metamorphism are shown in Tables 5.12a and 5.12b.

The granitic rocks of the Winisk Subprovince include granitic gneisses of indeterminate metamorphic grade (Bostock 1971) and clinopyroxene-bearing quartz monzonite and clinopyroxene-bearing mafic inclusions (Thurston et al. 1979, Appendix 1) perhaps indicative of amphibolite-facies metamorphism.

Suggestions that the Berens River Subprovince represents a deeper crustal level than greenstone belts in the adjacent Uchi and Sachigo subprovinces (e.g., Wilson 1971; Ayres 1978) are supported by limited geothermometric and geobarometric data (*see* Table 5.11) (Stone 1990). For example, the garnet-biotite thermometer (Ferry and Spear 1978) gives temperatures of 450° to 530°C for non-migmatized wackes in the Red Lake greenstone belt and in the southern Uchi Subprovince. Migmatites of the southern English River Subprovince attained temperatures of up to 650°C. In contrast, small units of metasedimentary migmatites in the southern and northern Berens River Subprovince record temperatures of 600° to 720°C; in most cases garnet core compositions give higher temperatures than rim compositions.

Metamorphic pressure determinations are scarce and highly variable, but they show that migmatites were buried deeper in the Berens River Subprovince than in the English River Subprovince. The plagioclase-garnet-aluminosilicate-quartz geobarometer (Koziol and Newton 1988) gives pressures of 810 to 425 MPa for the Berens River samples in contrast with 452 to 320 MPa for English River samples using garnet core compositions.

Felsic plutonic rocks in the Berens River Subprovince (Card and Ciesielski 1986) show a correlation between decreased aluminum content of amphiboles and decreasing relative age of the various granitic units (Stone 1990). The data imply that younger rocks were emplaced at higher levels, possibly due to progressive unroofing of the Berens River Subprovince during 2.7 billion-year-old plutonism.

Table 5.11. Geobarometry—Uchi, Sachigo and Berens River subprovinces.

		Comments	References
THERMOMETRY			
garnet–biotite*			Ferry and Speer 1978
Red Lake, Southern Uchi belt	450 to 530°C	single sample—inclusion in granitic unit 855°C	
Berens River	600 to 720°C		
Favourable Lake	597 to 721°C		
BAROMETRY			
plagioclase–garnet–aluminosilicate–quartz			Newton and Haselton 198
English River	320 to 339 MPa	—	
Berens River	810 to 425 MPa		
Favourable Lake	620 to 450 MPa		
pyroxene			
garnet–orthopyroxene			Ellis and Green 1980
Uchi Subprovince	5.80 to 7.40 MPa	—	
Favourable Lake	5.64 to 7.70 MPa		
hornblende			
Uchi Subprovince	0.84 MPa	—	Hammarstrom and Zen 1986
Berens River	2.07 to 5.4 MPa		Hollister et al. 1987
Favourable Lake	2.76 MPa		Blundy and Holland 1990

* *Temperatures listed are for cores of grains.*

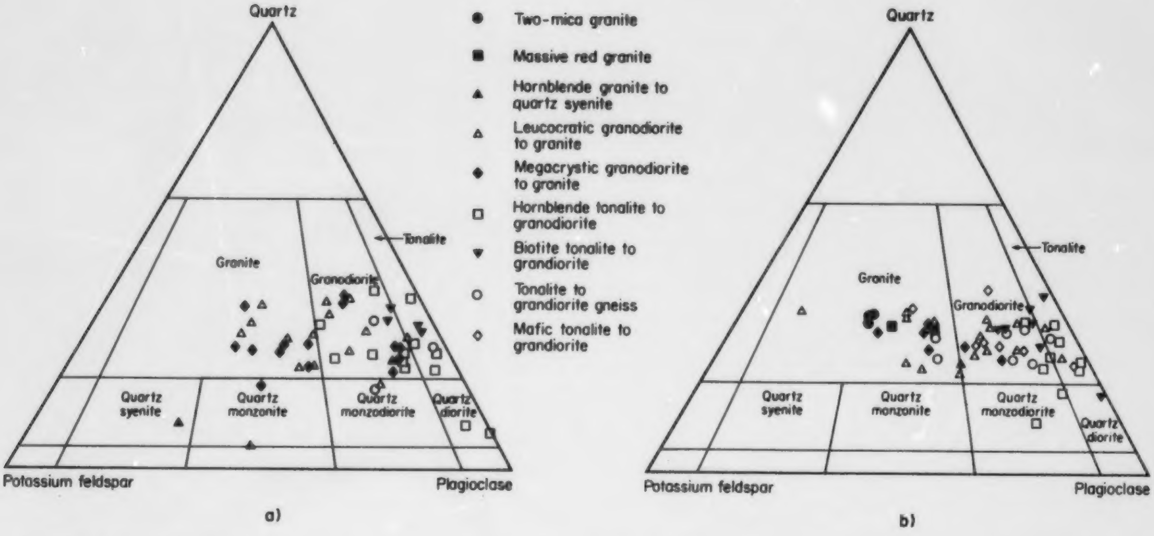


Figure 5.18. Ternary plots of modal quartz-plagioclase-potassium-feldspar of Berens River Subprovince granitoid rocks from **a)** Cobham Lake area, and **b)** Nungesser Lake area, *from* Stone (1990).

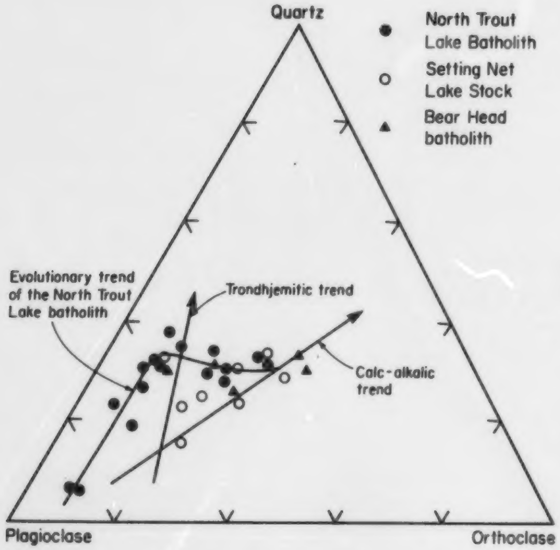


Figure 5.19. Variation in modal mineralogy in the North Trout Lake batholith *from* Ayres (1974), and the Setting Net Lake stock and Bear Head batholith (data *from* Stone 1990).

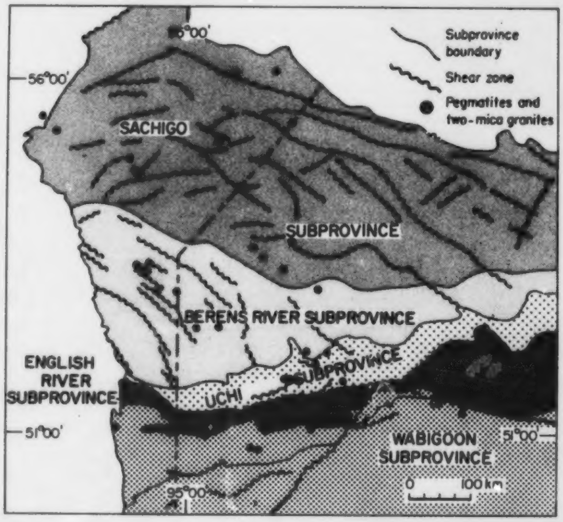

56°00'
56°00'
Subprovince boundary
Shear zone
Pegmatites and two-mica granites
SACHIGO
SUBPROVINCE
BERENS RIVER SUBPROVINCE
ENGLISH RIVER SUBPROVINCE
UCHI
SUBPROVINCE
51°00'
51°00'
WABIGOON SUBPROVINCE
95°00'
0
100 km

Figure 5.20. Distribution of fractionated pegmatites and two-mica granites in the Sachigo and Berens River subprovinces of Ontario and Manitoba, *modified from* Breaks and Osmani (1989).

Uranium-lead geochronology on metamorphic garnets show that the western part of the Sachigo Subprovince was subjected to an amphibolite-facies metamorphic event between 2744 and 2738 Ma, followed by 2 granulite-facies events between 2700 and 2687 Ma, and between 2646 and 2644 Ma (Mezger 1989). Based on sphene and zircon geochronology, the south-central part of the Sachigo Subprovince was metamorphosed between the age of the youngest widespread batholith phase at 2711 Ma, and the age of titanite in an amphibolite-facies clast (2679 Ma) (Corfu and Ayres, in press) in the North Trout assemblage of the Favourable Lake greenstone belt.

We conclude that metamorphism was largely brought about by the emplacement of posttectonic intrusive rocks. The regional variations within the northwestern Superior Province may indicate that the northwestern Superior Province, as presently exposed, represents an exhumed mid-crustal environment.

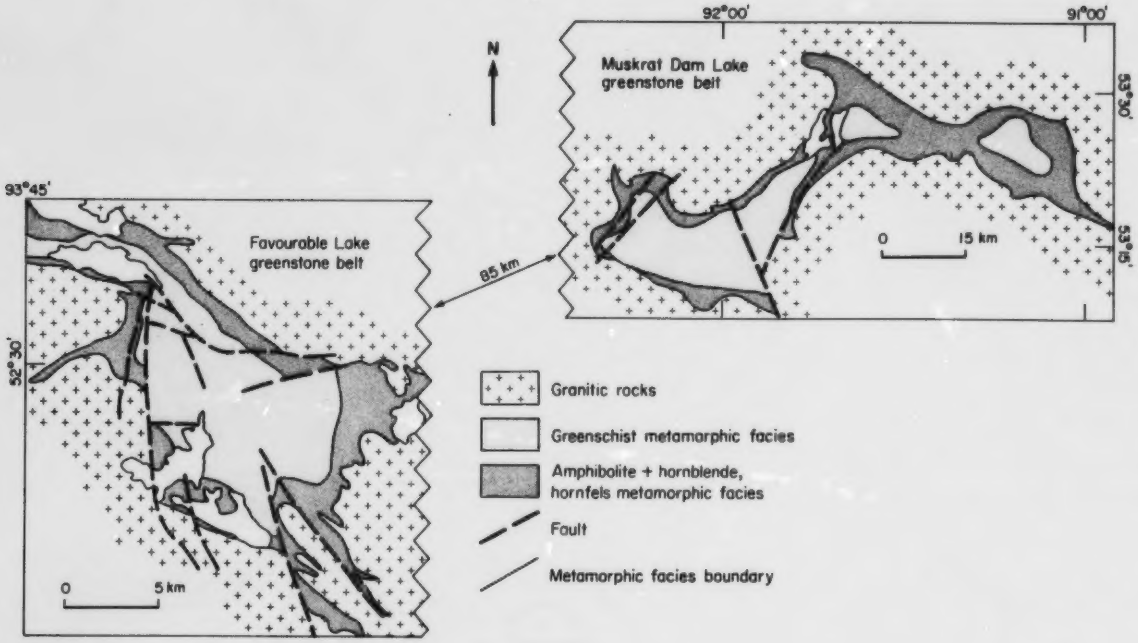

N
92°00'
91°00'
Muskrat Dam Lake greenstone belt
53°30'
53°15'
0 15 km
93°45'
Favourable Lake greenstone belt
85 km
52°30'
0 5 km
Granitic rocks
Greenschist metamorphic facies
Amphibolite + hornblende, hornfels metamorphic facies
Fault
Metamorphic facies boundary

Figure 5.21. Distribution of metamorphic zones in 2 typical greenstone belts of the Sachigo Subprovince (*modified from* Ayres 1978).

STRUCTURAL GEOLOGY

Structures in Greenstone Belts

The Berens River and Sachigo subprovinces are characterized by steeply dipping foliations and steeply plunging lineations in the rocks of the greenstone belts. Sequences of structural events have been developed for many of the greenstone belts (e.g., Breaks et al., in prep.; Cortis et al. 1988; Ayres 1974; Thurston et al. 1987) and appear to be similar. In general, the region is the product of multiple deformation events; our approach is to describe the events and the evidence for them through the use of selected examples. A later section will describe the effects of the structural history in terms of the structural styles observed. In a later section, structural development is integrated with geochronological results in a terrane analysis.

The earliest structures identified in some greenstone belts (e.g., Stowe 1974; Martyn 1987) are minor folds and more extensive thrusts in the metasedimentary assemblages that took place prior to lithification, penetrative deformation and lineation development. Early folding events are postulated for the North Caribou Lake greenstone belt (Breaks et al., in prep.), the Sandy Lake greenstone belt (Cortis et al. 1988), and the Big Trout Lake–Swan Lake greenstone belt (Hudec 1964). Evidence for these events is commonly found in a disparity in the expected sequence of absolute ages based upon stratigraphic criteria (e.g., Corfu and Stott 1989). Thrusting seems to have taken place along bedding-parallel shear zones that bear no consistent relationship to surrounding batholiths or to changes in metamorphic grade, and therefore occurred early in greenstone belt development. Sequences disrupted by thrust faults in the Sachigo and Berens River Subprovince greenstone belts may be indicated by the following:

1. the shear zones separating the assemblages of the Favourable Lake greenstone belt (Ayres 1988)
2. the repetition of ages of 2734 Ma (D.W. Davis, Royal Ontario Museum, personal communication, 1989) within the Muskrat Dam assemblage of the Muskrat Dam Lake greenstone belt
3. the six-fold repetition of a stratigraphic sequence at Oxford Lake (Hubregtse 1978)
4. the inferred fault, parallel to the upper contact of the Keeyask assemblage in the North Caribou greenstone belt
5. the "front-to-front" relationship between the Northern Horseshoe metasedimentary assemblage and the Horseshoe metasedimentary assemblage
6. the "back-to-back" relationship between assemblages in the Hornby Lake greenstone belt

Early deformation has been postulated to explain isoclinally refolded folds along the limbs of D_2 folds in the North Caribou greenstone belt (Breaks et al. 1986).

The second deformation event produced upright, belt-parallel, shallow to steeply plunging folds in Sachigo and Berens River Subprovince greenstone belts. Examples of these structures include the major belt-parallel folds that bear a strong spatial relationship to the neighbouring granitic batholiths, for example, those found at North Caribou Lake (Breaks et al., in prep.), Horseshoe Lake (Jensen 1987,

1988) and North Spirit Lake (Wood 1977, 1980b). Examples of early folds refolded during the D_2 event are seen in iron formation on Opapimiskan Lake in the North Caribou greenstone belt (Breaks et al. 1986)(Figure 5.22). Foliations developed at the time of D_2 folding are overprinted by contact metamorphic aureoles around individual granitic batholiths (cf. Breaks et al., in prep.; Ayres 1978). The axes of major folds are commonly warped. The spatial relationship of the warps indicates their most likely cause was the emplacement of regional batholiths such as at North Caribou Lake (Breaks et al., in prep.), Sandy Lake (Cortis et al. 1988), Muskrat Dam Lake (Ayres 1969) and Lingman Lake (Wilson 1987). In addition to folding, the D_2 event also produced shear zones parallel to D_2 fold axes and lithologic contacts. These shear zones are distinguishable from D_1 shear zones only where the 2 fold and shear systems are not parallel.

Examples of D_2 shear zones include those separating assemblages in the Muskrat Dam Lake greenstone belt (Thurston et al. 1987), the shear zone extending the whole length of Sandy Lake (Thurston et al. 1987; Cortis 1988), and the minor, unnamed, shear zones between assemblages in the North Spirit Lake greenstone belt.

The third stage of deformation in greenstone belts produced open to gentle, probably upright, asymmetric folds and chevron crenulation of D_2 foliations and associated axial planar cleavages in the North Caribou greenstone belt. Macroscopic D_3 folds may be the cause of changes in the trend and plunge of D_2 lineations, particularly in the southeastern part of this greenstone belt.

A series of major shear zones within the Berens River and Sachigo subprovinces postdate the major folding events, in that they transect the regional-scale folds (Osmani and Stott 1988). These shear zones host spatially associated sheets of muscovite-bearing granite. As these types of melts are developed in metamorphic conditions greater than those that have been imposed on the surrounding rocks, this relationship suggests that the late regional-scale shear zones may be deep-seated structures (Stott and Corfu, this volume). These shear zones of uncertain age will be discussed in the Terrane Analysis section.

Relationship of Greenstone Structures to Granitoids

Multiple generations of felsic plutonic rocks have intruded the interiors and margins of greenstone belts in the northwestern Superior Province. For example, the North Caribou greenstone belt was cut by a tonalite pluton at 2.99 Ga and a crescentic tonalite pluton was emplaced along margins of the belt at 2.87 Ga. Sodic and potassic plutons intruded the Favourable Lake greenstone belt at 2.73 and 2.71 Ga (Corfu and Ayres, in press). Although the gneissic tonalite suite could be basement to some greenstone belts (Hillary and Ayres 1980), emplacement of younger granitic suites appears to have deformed not only pre-existing greenstone belts, but plutonic rocks as well. For example, many units of tonalite share a common belt-like shape with greenstone units.

Foliation trends in the gneissic tonalite suite are similar to those in adjacent greenstone belts: they diverge around younger batholiths and are generally parallel to belt boundaries. In contrast, plutons of the diorite-monzonite-granodiorite suite and the massive granodiorite to granite suite display foliation trajectories that are concentrically oval and best developed at the pluton margins, such as south of Muskrat Dam Lake (*see* Tectonic Assemblages of Ontario, Maps 2575 to 2577, map case) and the North Caribou batholith (Breaks et al., in prep.).

The young oval granite bodies are largely undeformed at their centres. Two explanations may be offered:

1. The less deformed centres of the batholiths, as well as geochronologic data, indicate that pre-existing country rocks were constricted by young felsic intrusions. The oldest units (greenstone belts and gneisses) have been deformed by successive intrusions of sodic and potassic plutons and have undergone shortening which caused the strong alignment or realignment of foliations parallel to greenstone belt margins. Major structures such as D_2 folds in greenstone belts could have been generated by this deformation process. Although the margins of oval granite batholiths are slightly deformed, the oval batholiths have never been intruded by younger bodies, and hence retain their original shapes and fabrics.
2. The cores of plutons may have been protected from further deformation by strain softening at the pluton margins, which may have absorbed most of the strain.

Both greenstones and granites responded to the latest deformation events related to the development of shear zones such as the Bear Head fault zone. Although the Bear Head fault zone may be as old as 2769 million years, based on the age of a sheared monzonite (Corfu and Ayres 1984), a 2696 to 2697 million-year-old granodiorite is also locally deformed (Corfu and Ayres 1984), suggesting that the Bear Head fault had a movement history that outlasted plutonism.

Structural Style

The structural style of the greenstone belts of the northwestern Superior Province is dominated by homoclinal panels that are bounded by shear zones (Table 5.13) .

By analogy with most other greenstone successions in the Superior Province (cf. Jackson and Fyon, this volume; Blackburn et al., this volume), this style is consistent with the repetition of stratigraphic units by thrusting or the superposition of supracrustal panels of a variety of ages. Folded panels are present on a small scale in a number of belts (*see* Table 5.8). Examples include the Favourable Lake greenstone belt and the North Spirit Lake greenstone belt that were assembled by thrusting at about 2710 Ma (Corfu and Ayres, in press). The Hayes River Group contains 6 mafic to felsic volcanic cycles at southern Knee Lake in Manitoba (Hubregtse 1978) that, by analogy with similar situations in southern Superior Province (Noble et al. 1989; Beakhouse 1988; Jackson and Fyon, this volume), may represent structural panels.

Table 5.12a. Inter–relationship of metamorphism and tectonism in the North Caribou greenstone belt (after Breaks et al. 1990).

Lithology	Deformation Event	Metamorphism	Age (Ma)	Comments	References
diabase	Brittle deformation in major fault zones		2827	Lower age limit of faulting provided by age of galena in carbonate cementation of fault breccia	R. Thorpe, GSC, personal communication to F. Breaks 1990
peraluminous granite suite		Autometasomatized	Age uncertain; may be similar to 2850 to 2860 gold mineralization at Musselwhite property	Extensively albitized and locally greisenized; late and posttectonic dikes	Breaks et al. (in prep.)
	D_3	Local S_3 cleavage with chlorite, muscovite; late grunerite in ironstone			R. Thorpe, GSC, personal communication to F. Breaks 1990
Zn-P-Ag mineralization in Castor–Pollux Lakes deformation zone			2852		R. Thorpe, GSC, personal communication to F. Breaks 1990
Schade Lake tonalite gneiss	Late mylonite zones localized along greenstone belt boundaries		2856.0 ± 1.8		D.W. Davis, ROM, unpublished U-Pb zircon age
North Caribou batholith (NCB)		Contact metamorphic amphibolite aureole overprinted upon $D_2 - M_2$ fabric on mafic metavolcanics	2869 ± 3.6		Breaks et al. (in prep.) D.W. Davis, ROM, unpublished U-Pb zircon age
	D_2 Major ductile deformation; development of F_2, S_2 fabric and meso- and macroscopic ductile fault zones	M_2 Sinistral rotation of M_{1A} porphyroblasts and reworking of S_1 fabric	D_1 and D_2 developed between 2871 and 2958.7		Breaks et al. (in prep.)
		M_{1A} Development of garnet, andalusite and cordierite porphyroblasts in ES			Breaks et al. (in prep.)
	Mainly preserved as internal planar fabric in garnet, andalusite and cordierite; M_{1A} porphyroblasts and D_1 isoclinal folds refolded by D_2	Planar fabric in Eyapamikama metasediments (ES) composed of chlorite, biotite and muscovite			
Eyapamikama metasediments (ES); second cycle sedimentation (alluvial and submarine fan turbidite basin infill)			2871 to 2958.7 ± 1.8	Age of sedimentation bracketed between age of tonalite clast in conglomerate and macroscopic warping of ES by 2871 NCB	Breaks et al. (in prep.) D.W. Davis, unpublished U-Pb zircon ages

Table 5.12a. Inter-relationship of metamorphism and tectonism in the North Caribou greenstone belt (after Breaks et al. 1990).

Lithology	Deformation Event	Metamorphism	Age (Ma)	Comments	References
Unconformity with Keeyask Lake Metavolcanics and possibly on NCB					
North Rim (NRV) and South Rim (SRV) metavolcanics (Fe tholeiites, minor komatiites)			2932± 3 with older, possibly detrital 2965± 2 zircons		D.W. Davis, ROM, unpublished U–Pb zircon age
Keeyask metavolcanics (komatiites, Mg tholeiites associated with rifting)				Relation to NRV and SRV unknown	
Keeyask metasediments; first cycle sedimentation in stable shelf platform environment			Younger than 2978± 3 and plausibly older than NRV at 2932± 3		
Unconformity					
Agutua Arm (Weagamow Lake) metavolcanics (AAV); arc volcanics, Fe tholeiites dominate			2978.0± 1.8		D.W. Davis, ROM, unpublished U-Pb zircon age
Weagamow batholith (plausible basement to AAV)			2990.0± 1.8		D.W. Davis, ROM, unpublished U-Pb zircon age

Table 5.12b. Development of the Favourable Lake area. (events in order of youngest to oldest) (after Ayres and Corfu, in press)

Age	Event
2711 Ma	isoclinal folding, batholith emplacement, metamorphism
2725 to 2711 Ma	stacking of thrust slices
2725 Ma	development of the North Trout assemblage
2734 Ma	development of the Northwind assemblage
2870 to 2858 Ma	development of the Eastern Trout and South Trout assemblages
2925 to 2870 Ma	deposition of the upper part of the Setting Net assemblage after submergence
2925 Ma	deposition of lower part of the Setting Net assemblage

Table 5.13. Structural styles in the northwestern Superior Province.

Style	Example	Figure Number	References
1) homoclinal panels			
a) greenstone belt scale	Sandy Lake	11	Thurston et al. 1987
	Upper Windigo Lake	5b	Cortis et al. 1988
b) Assemblage scale	Nekence assemblage	9	Ayres 1969
	all assemblages in: Muskrat Dam Lake greenstone belt	9	Breaks et al., in prep.
	North Caribou greenstone belt	3	Thurston et al. 1979
	Wunnummin Lake greenstone belt	8	
2) folded panels			
a) assemblage scale	Eyapamikama assemblage, North Caribou greenstone belt;	3	Breaks et al., in prep.
	Hewitt assemblage, North Spirit greenstone belt;	7	
	Setting Net, Northwind assemblages, Favourable Lake greenstone belt	6a	Ayres 1988
3) thrust–banded panels	Favourable Lake greenstone belt	6a	Ayres 1988
	Swan Lake–Big Trout Lake greenstone belt (?)	8	this report
4) back to back	Western McInnes versus Southern McInnes assemblages, McInnes Lake greenstone belt	15	Cortis et al. 1988
5) front to front	a) Northern Horseshoe versus Central Horseshoe assemblages, Horseshoe Lake greenstone belt	5a	Jensen 1987
	b) Northwind versus North Trout assemblages, Favourable Lake greenstone belt	6a	Ayres 1988
6) outward facing off granitic dome	2.9 Ga platformal sequences, Sachigo Subprovince—*see* text	—	this report

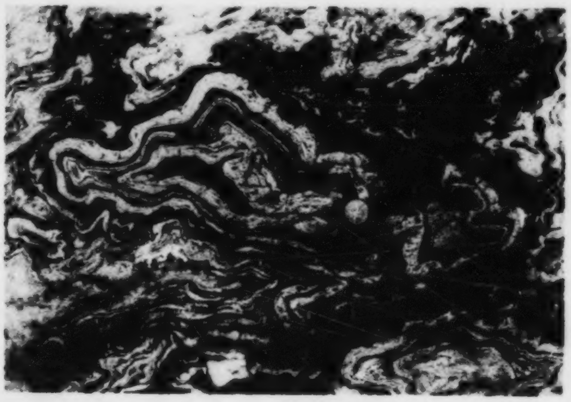


Figure 5.22. Photograph of dome and basin interference structures in the iron formation at Opapimiskan Lake (*from* Breaks et al. 1986). The coin is about 2 cm in diameter.

The structural style relationships portrayed in Figures 5.3, 5.5a, 5.6a, 5.7 and 5.11 suggest that, in some cases, greenstone assemblages were formed into the belts as we now know them prior to their interaction on a terrane or subprovincial scale. For example, the arc metavolcanic rocks of the 2980 million-year-old Agutua Arm assemblage is unconformably overlain by the approximately 2980 million-year-old Keeyask assemblage platform, and is in tectonic contact with the 2932 million-year-old North Rim component of the McGruer assemblage. The 2871 million-year-old North Caribou batholith imposes a metamorphic and strain aureole upon the belt. Therefore, assembly of the belt must have taken place between 2980 Ma and 2870 Ma.

The structural style of the northwestern Superior Province may be a function of granite emplacement processes, terrane interaction, or both. The presence of rocks with widely varying ages within single greenstone belts (e.g., Favourable Lake greenstone belt, North Spirit Lake greenstone belt) suggests that the processes of assemblage and terrane interaction and tectonic transport are significant in the development of structural style. Emplacement of granitic rocks, especially sheet-like granitic bodies, has also influenced the form and proportion of greenstones seen in the northwestern Superior Province.

Geophysical Expression of Major Structures

Major structures within the northwestern Superior Province are apparent on small-scale gravity and magnetic maps. The continental scale gravity map (CGAMNA 1987) shows regional-scale gravity highs over the following areas: 1) the Berens River Subprovince, indicative perhaps of the less than usual depth to dense granulite-facies rocks; and 2) the Muskrat Dam Lake and Sandy Lake greenstone belts,

Table 5.14. Metallogenic associations—northwestern Superior Province.

Environment	Commodity	Deposit Description	References
platform	Au, U	detrital grains in quartz arenites, 1) detrital Au found at North Caribou Lake, and 2) U minerals in quartz arenites	Thurston and Chivers 1990, Roscoe 1990
	Fe	iron formation in upper part of platform sequences, e.g., North Spirit Lake	Wood et al. 1986
	Ni	could be in komatiites of platform sequences	
arc volcanism	Cu, Zn, Pb	volcanogenic massive sulphide deposits—none known in northwestern Superior Province	
	Pb, Zn, Ag, Au	epithermal (?) vein, deposit in caldera-fill sequence of formation H, Favourable Lake	Adams 1976
	Cu	disseminated chalcopyrite, pyrite in altered mafic flows beneath an impermeable cap	L.D. Ayres, personal communication
	Pb, Ag, Au	pyrrhotite, pyrite, argentite, galena, sphalerite, arsenopyrite in veins cutting wackes associated with mafic flows at Borland Lake	Stone 1989
	PGE, Cr	associated with synvolcanic(?) ultramafic-mafic sill at Big Trout Lake	Macdonald and Cherry 1988
arc plutonism	U	small pegmatite-related U deposits associated with quartz monzonite to granite in Berens River Subprovince at Hornby Lake and Favourable Lake	Breaks et al. 1978
	Cu, Mo	porphyry-style mineralization in Setting Net Lake stock in the Favourable Lake greenstone belt	
late unconformable basins and shear zones	rare metals, Au	pegmatites and Au-anomalous two-mica granites are spatially associated with major shear zones	Breaks et al. 1986, Stone 1989
	Au	discordant vein systems in various rock types spatially associated with D_2 shear zones and major late shear zones	

probably indicative of the overall size of the area underlain by relatively dense basaltic rocks. The gravity field shows a discontinuity coincident with the Ponask Lake shear zone and the Bear Head fault zone (*see* Figure 5.2), suggesting that these are major structures. The 1:1 000 000-scale magnetic maps accompanying this volume (*see* Maps 2584 and 2588, map case) clearly shows foliation trends within the gneissic tonalite suite and the foliated tonalite suite throughout the northwestern Superior Province. Likewise, the maps show a contrast in the texture of the magnetic pattern between the various assemblages making up most greenstone belts. Throughout this part of the Superior Province, curved fabric relationships are demonstrably associated with major shear zones (cf. Osmani and Stott 1988).

MINERAL DEPOSITS

An exhaustive catalogue of northwestern Superior Province mineral deposits is inappropriate here; instead we comment on the spatial and temporal patterns of mineral deposit distribution in this region. Mineral deposit locations are shown on 1:253 440-scale compilation maps published by the Ontario Geological Survey. The major deposit types in this part of the Superior Province are listed in Table 5.14. Metallogenic trends are described in the chapter by Fyon et al. (this volume) drawing examples from throughout Ontario.

Quartz-rich Archean platformal metasedimentary rocks have only recently been discovered in the Superior Province (Wood et al. 1986). The Sachigo Subprovince contains the most extensive areas of this assemblage type. The unique metallogenic character of these quartz-rich metasedimentary rocks has been described only recently (Groves and Batt 1984; Thurston and Chivers 1990).

Slightly anomalous scintillometer readings have been obtained from Archean quartz arenites in the Slave Province and in the Keeyask assemblage (S. Roscoe, Geological Survey of Canada, personal communication, 1990), suggesting some uranium potential in these rocks. The iron deposits in the Nemakwis assemblage at North Spirit Lake are a good example of iron mineralization in platform sequences within the northwestern Superior Province. Although barium and tungsten deposits have been recognized in other platformal assemblages, principally in the Pilbara block of Western Australia (Groves and Batt 1984), none have been discovered in the Sachigo Subprovince.

Some features of arc magmatism are well displayed in the greenstone belts of the Sachigo Subprovince. The Setting Net Lake porphyry-style copper-molybdenum deposit (Nunes and Ayres 1982; Ayres et al. 1982) is one of only a very few Archean deposits of this type. The porphyry-style mineralization and the epithermal mineralization in the 2926 million-year-old caldera at Favourable Lake (Buck 1978) preserve areas of high crustal level rocks and mineral deposits relative to other parts of the Superior Province, a result of minimal crustal thickening and/or attendant erosion. Volcanogenic massive sulphide deposits have not yet been found in the 2.7 billion-year-old arc volcanic sequences of the northwestern Superior Province. However, the total volume of 2.7 billion-yearold metavolcanic rocks in the area is quite minor. It is interesting to note that no volcanogenic massive sulphide deposits are associated with any of the 2.8 to 2.9 billion-year-old metavolcanic assemblages.

Four types of mineral deposits appear to be spatially associated with major faults in the region: vein silver and gold, rare-metal granites and pegmatites, lode gold deposits, and minor uranium mineralization. Deposits of silver and minor gold (Adams 1976; Stone 1989) occur in veins cutting the stratigraphy at the Berens River Mine (Adams 1976) within the 2926 million-year-old caldera sequence at Favourable Lake. The metal ratios suggest an epithermal affinity, but the potassic metasomatism typical of this deposit type is not documented. Silver-rich veins occur at Borland Lake where 391 985 tons of ore, grading 289 g/ton Ag and 0.69 g/ton Au, occur along the interface between the Bear Head fault zone (Aerodat Ltd. 1985). Other less significant occurrences are summarized by Stone (1989). Breaks and Osmani (1989) have described a close spatial association between two-mica granites and related fractionated pegmatites with associated rare-metal mineralization, and the regional-scale shear zones of Osmani and Stott (1988). Rare-metal pegmatite deposits associated with two-mica granites situated along major shear zones have been dated at about 2653 Ma (Mezger et al. 1990). Most occurrences in the northern Superior Province are within Manitoba (Cerny and Trueman 1978). Several small, low-grade uranium occurrences were documented by Bond and Breaks (1978) along the same fault. Breaks et al. (1986) described an association of anomalous gold background values with fluorite-bearing two-mica granites occurring along major shear zones within the North Caribou greenstone belt.

Lode gold deposits in the northwestern Superior Province are spatially associated with regional-scale shear zones (Colvine et al. 1988; Piroschco and Shields 1985; Osmani and Stott 1988). The principal deposits in the northwestern Superior Province are: 1) the Musselwhite deposit in the North Caribou greenstone belt, with more than 1 000 000 tons grading approximately 0.20 ounce Au per ton (*The Northern Miner*, March 5, 1981); 2) the Lingman Lake Mine, with reserves of about 574 000 tons (Wilson 1987); 3) the Berens River Mine, which produced 157 341 ounces Au; and 4) the Sachigo River Exploration Co. Ltd. mine, which produced 46 457 ounces Au (Ayres et al. 1969).

Sulphides from lode gold deposits in the Sachigo Subprovince have yielded model Pb ages clustered around 2850 Ma (R. Thorpe, Geological Survey of Canada, personal communication, 1990). Results from the Arseno Lake gold occurrence yielded ages of 2853 to 2850 Ma, and the North Caribou River silver occurrence on Agutua Arm yielded a model age of 2827 Ma. Lead isotopic analysis of sulphides from gold deposits provides an indication of the age of the source rocks for the mineralization. Volcanism within the North Caribou greenstone belt occurred between 2990 and 2932 Ma (*see* Table 5.1) and posttectonic plutonism took place 2869 Ma (D.W. Davis, Royal Ontario Museum, personal communication, 1990). The Pb isotopic results therefore indicate that source rocks for these deposits were slightly younger than posttectonic plutonism. In the southern Superior Province, lode gold mineralization was dated at 2682 Ma by analysis of hydrothermal zircons using the U-Pb method (Claoue-Long et al. 1990), and Pb isotopic results for Abitibi Subprovince gold deposits cluster at about 2.6 to 2.7 Ga. Therefore, the Sachigo Subprovince results indicate the existence of extensive crust that predates 2.7 Ga. The results may also provide a tentative indication of the existence of a lode gold mineralizing event before 2.7 Ga. This suggestion will be the basis of further work in the northwestern Superior Province.

POST–ARCHEAN EVENTS

The northwestern part of the Superior Province was the site of minor intracratonic magmatism during the Proterozoic. Alkalic stocks occur at Wapikopa Lake (2534 Ma), Schryburt Lake (1145 Ma), Big Beaver House (1109 Ma) and Carb Lake (1826 Ma). These intrusions are described more fully in Sage (this volume). No large-scale tectonic controls on the locations of these stocks have been suggested. However, we note that all alkalic stocks in the northwestern Superior Province are spatially relatable to major shear zones. Three diabase dike sets cut the northwestern Superior Province. They are the 2171 million-year-old Marathon swarm, the 1888 million-year-old Molson swarm and the 1267 million-year-old Mackenzie swarm. The extent, character, and large-scale controls on these swarms are described by Osmani (this volume).

Ermanovics and Wanless (1983) have observed a systematic northward and westward younging of K-Ar ages through the northwestern Superior Province toward the Trans–Hudson Orogen. The ages exhibit a near-systematic decrease toward the 1.8 Ga age of that orogen, providing evidence for a Proterozoic thermal event affecting the northwestern Superior Province.

TERRANE ANALYSIS

The Sachigo and Berens River subprovinces were the major divisions of the northwestern Superior Province when compilation of this volume and the accompanying maps commenced (Card and Ciesielski 1986). Therefore, this chapter consists of descriptions of the geographic area covered by these 2 subprovinces. Subprovinces of the southern Superior Province have been considered comparable to terranes, superterranes, or composite terranes (e.g., Hoffman 1989), displaying a relatively limited range of ages, coherent lithostratigraphic packages, and similar structural histories. The northwestern Superior Province displays a number of characteristics suggesting that it is not comparable to other Superior Province granite-greenstone subprovinces and ought not to be considered as a single terrane or a simple composite terrane (Sachigo and Berens River) based on the following points:

1. Duration of geological processes:
   The Sachigo Subprovince displays unusual durations of various geologic processes: posttectonic plutonism ranges from 2871 Ma (D.W. Davis, Royal Ontario Museum, personal communication, 1989) to 2711 Ma (Corfu et al. 1985), whereas the emplacement of the Sierra Nevada batholith spans a time range of 60 to 80 million years (Windley 1984, p. 284). Volcanism spans a range from 3023 Ma (Corfu and Wood 1986) to 2734 Ma (Corfu and Ayres, in press). The age of shear zone activity within the subprovince ranges from 2770 Ma (Corfu and Ayres 1984) to approximately 2650 Ma (Mezger 1989). Metamorphic events range from 2925 Ma (Breaks et al., in prep.) to the 2679 Ma sphene age of an amphibolite-facies clast in metavolcanic rocks of the Favourable Lake greenstone belt (Corfu and Ayres, in press).
2. Variety of depositional environments:
   Sachigo and Berens River subprovince greenstone belts are divisible into platformal, mafic plain, and pull-apart basin lithostratigraphic associations (Thurston and Chivers 1990) exhibiting large ranges in age.
3. Presence of shear-zone–bounded blocks (Osmani and Stott 1988):
   The blocks were defined on the basis of detailed mapping, helicopter-supported reconnaissance, geochronology, field checking, and aeromagnetic and airborne gamma ray spectrometer surveys as reported by Osmani and Stott (1988). The shear zones define blocks within the northwestern Superior Province which have distinct lithostratigraphic associations representing varying depositional environments (Thurston and Chivers 1990). The shear-zone–bounded blocks display limited age ranges for granitic and supracrustal rocks relative to the northern Superior Province as a whole (*see* geochronological data in Tables 5.1, 5.3 to 5.7). Individual blocks are defined on the basis of the above parameters in descriptions of individual terranes in the following pages.

Terrane analysis must be clearly understood to be a purely descriptive tool which emphasizes geological similarities of discrete blocks, and which is not directly dependent upon any particular genetic hypothesis. However, plate tectonic processes represent a favoured explanation for the fundamental meaning of terranes as descrete blocks with distinctive geological histories and how they came to be assembled into their present configuration. We suggest here that the northwestern Superior Province is composed of at least 4 terranes. These terranes are comparable in rock types, age ranges, and structural character to subprovinces in the southern part of the Superior Province, albeit of smaller size (cf. Hoffman 1989; Percival and Williams 1989; Card 1990; Williams 1990).

In a seminal paper, Langford and Morin (1976) explained granite-greenstone subprovinces in southern Superior Province as the product of arc-arc accretion. Krogh and Davis (1971) provided early documentation of the southward younging of the age of volcanism. More recent work has compared the metasedimentary Quetico Subprovince to modern forearc accretionary prisms (Percival and Williams 1989; Williams 1990). Volcanologic, petrographic, and geochemical evidence has permitted discrimination of continental and oceanic volcanic arcs (Jackson and Sutcliffe 1990). Thurston and Chivers (1990) have shown that granite-greenstone subprovinces comprise diverse lithostratigraphic associations representing platforms, arcs, mafic plain volcanism (oceanic?) and pull-apart basins. Wyman and Kerrich (1988) have shown that the geochemistry of late Archean magmatism is comparable to the late stages of volcanic arcs. There is a growing consensus (e.g., Langford and Morin 1976; Hoffman 1989; Percival and Williams 1989; Card 1990; Williams 1990) that the southern Superior Province in particular was assembled by plate tectonic processes.

To define terranes in the northwestern Superior Province, constraints on the ages of shear zones forming terrane boundaries are desirable. As few direct determinations are available, indirect methods involving comparison with the southern Superior Province are employed. Corfu and Stott (1986) have shown that in the southern Superior Province, Timiskaming-type or late metavolcanic-metasedimentary sequences in greenstone belts are closely related temporally to the docking of the Quetico and Wawa subprovinces. Similar geological processes are thought to occur at the Uchi–English River (Thurston and Stott 1989) and Wabigoon–Quetico Subprovince boundaries (Davis et al. 1989). In the light of these relationships, the age of docking of terranes in the northwestern Superior Province may be approximated by the age of pull-apart basin assemblages.

The following terrane analysis includes the Sachigo, Berens River, and the northern part of the Uchi Subprovince, described by Card and Ciesielski (1986). The distribution of greenstone ages (Figure 5.23) requires a tectonic model involving one of 3 options:

1. ensialic rifting of 2.9 and 2.8 billion-year-old cratonic masses to permit development of 2.7 billion-year-old greenstones in locations well removed from known 2.7

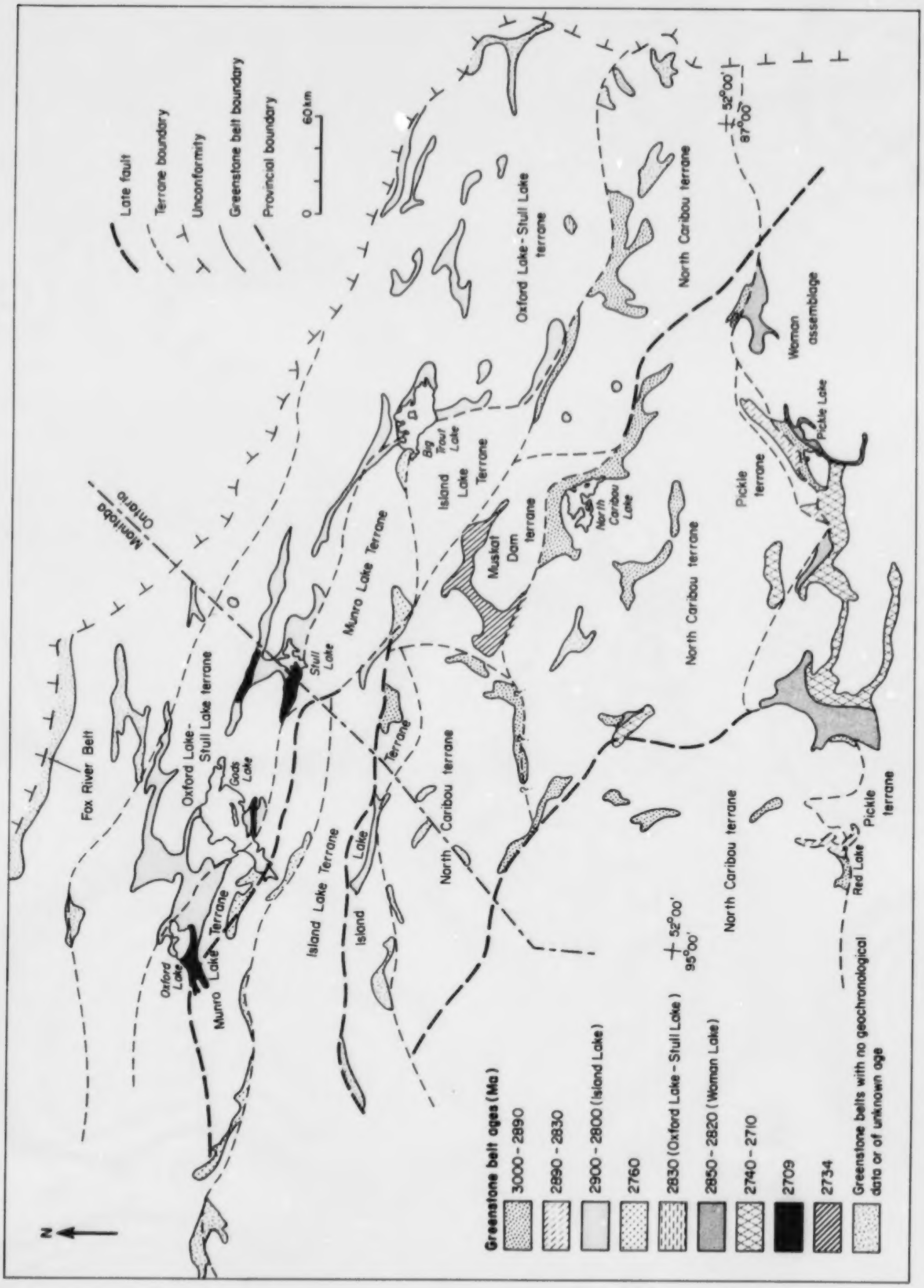

N
Late fault
Terrane boundary
Unconformity
Greenstone belt boundary
Provincial boundary
0
60 km
Greenstone belt ages (Ma)
3000 - 2890
2890 - 2830
2900 - 2800 (Island Lake)
2760
2830 (Oxford Lake - Stull Lake)
2850 - 2820 (Woman Lake)
2740 - 2710
2709
2734
Greenstone belts with no geochronological data or of unknown age
Fox River Belt
Manitoba
Ontario
Oxford Lake-Stull Lake terrane
Gods Lake
Oxford Lake
Munro Lake Terrane
Island Lake Terrane
Island Lake Terrane
North Caribou terrane
Stull Lake
Munro Lake Terrane
Muskrat Dam terrane
Island Lake Terrane
Big Trout Lake
North Caribou Lake
North Caribou terrane
Oxford Lake-Stull Lake terrane
North Caribou terrane
Pickle terrane
Pickle Lake
Woman assemblage
North Caribou terrane
Red Lake
Pickle terrane
52°00'
95°00'
52°00'
87°00'

Figure 5.23. Terranes of the northwestern Superior Province based on integration of geochronologic data, depositional setting of supracrustal rocks and the distribution of shear zones.

billion-year-old arcs of the southern Superior Province (Thurston and Stott 1989)

2. long distance tectonic transport of younger greenstones from arcs to the south
3. accretion of younger terranes between the blocks of older crust represented by the Berens River Subprovince and the Sachigo Subprovince.

None of these is particularly attractive, so we favour a new subdivision of the northwestern Superior Province, based on an accretionary hypothesis.

In this terrane analysis, the knowledge of the extent of regional-scale shear zones identified by Osmani and Stott (1988) was increased through inspection of contoured and coloured aeromagnetic data and compilation of regional-scale maps (*see* Shaded Image of Total Magnetic Field of Ontario, Maps 2584 to 2586, map case; MEM 1987).

Figure 5.23 shows the results of the terrane analysis based on:

1. the extended versions of the shear zones found by Osmani and Stott (1988)
2. a full compilation of U-Pb zircon ages throughout the Sachigo and Berens River subprovinces
3. assessment of the depositional environment of greenstone sequences based on the criteria of Thurston and Chivers (1990)
4. the age of cratonization of each terrane based on the age of posttectonic granitic units.

Inspection of Figure 5.23 reveals that, in common with the southern Superior Province, shear-zone–bounded blocks in the northwestern Superior Province are characterized by differences in structural style, structural trends, lithostratigraphic associations, and ages of volcanism and plutonism. The distinctive style and scale of each of these shear-zone–bounded blocks is comparable to Phanerozoic terranes (cf. Coney et al. 1980); therefore, we consider these blocks to be terranes or composite terranes. In a subsequent section, we review each of the proposed terranes, summarizing the distinctive features of each.

This synthesis is presented at an early stage in the geochronological examination of the northwestern Superior Province and emphasizes major subdivisions. Subsequent investigations will undoubtedly modify this preliminary subdivision, but it does provide a framework for future research.

Four lithostratigraphic associations, representing distinct depositional environments within Superior Province greenstone belts were identified by Thurston and Chivers (1990):

1. An early (approximately 3 billion-year-old) supracratonic platform and association of basal shallow-water quartz arenites and carbonates succeeded by argillites, ironstone and mafic to ultramafic flows, or mafic to ultramafic flows associated with quartz arenites and quartzofeldspathic metasedimentary rocks.
2. A mafic plain association composed of basal massive mafic flows with thin pillowed tops in edifices with low dips, intercalated with deep-water sulphidic argillites representing deep-marine central volcanic complexes.
3. Mafic to felsic cyclical volcanic assemblages representing arc volcanism.
4. Relatively young (ca. 2.68 Ga) pull-apart basin assemblages consisting of alluvial-fluvial sedimentary rocks and alkalic to calc-alkalic metavolcanic rocks deposited unconformably above older greenstones in fault-bounded basins.

All geochronological data cited are high- precision U-Pb ages on zircons unless otherwise noted. Table 5.15 summarizes the ages of cratonization, ages of potential basement rock types, and constraints on the ages of docking with adjacent terranes. The names of individual greenstone belts are taken from the relevant government reports and have no stratigraphic significance in the formal sense. Shear zone nomenclature is based on Osmani and Stott (1988) where possible.

Terrane boundaries were established along shear zones separating areas of differing ages, lithostratigraphic associations, structural styles and trends, and geochronology as documented in Table 5.15. Ages of inferred basement were derived from the presence of deformed granitoid units older than supracrustal units within terranes. No instances of true tectonic basement, such as a greenstone belt observed to be underlain by granitoid units, have been observed. Ages of quartz-rich metasedimentary rocks quoted in Table 5.15 are, almost without exception, ages derived from studies of detrital zircons and reflect source ages and provide some constraints upon depositional age. The ages of cratonization are based upon the ages of posttectonic plutons within the terranes. The docking of one terrane against another is derived from one or more of the following:

1. the age of stitching plutons
2. the age of pull-apart basins
3. the age of the latest metamorphism

North Caribou Terrane

This terrane is characterized by approximately 2.9 billion-year-old quartz-rich metasedimentary sequences of sialic provenance (D.W. Davis, Royal Ontario Museum, personal communication, 1989; Cortis et al. 1989) and carbonate platform-type greenstones (Thurston and Chivers 1990) commonly deposited above older arc sequences. The boundaries of the terrane and the types of supracrustal sequences within the greenstone belts are listed in Table 5.15. Gneissic tonalitic rocks which are 2.9 to 3.0 billion years old in the Lake Winnipeg area (Krogh et al. 1974) represent possible basement within the terrane. The cratonization age of this terrane is tentatively identified as 2.87 Ga, the age of the North Caribou batholith (Stott et al. 1989).

The terrane consists of the 2.9 billion-year-old parts of the Red Lake and the Confederation Lake greenstone belts (part of the Uchi Subprovince as formerly designated), the Berens River Subprovince, and the southern part of the

Table 5.15. Terranes of the northwestern Superior Province.

Terrane	Boundaries	Greenstone Belts, Assemblage Types, Ages (Ma)	Structural Trends	Cratonization Age (Ma)	Docking Age (Ma)	References
North Caribou terrane						
northern subunit	N: Shear zone south of Island Lake Rottenfish, Shear Zone S: Sandy Lake, shear zone E: Rottenfish Shear Zone W: Bear Head fault zone	1) Southern Island arc(?) 2) Sandy Lake North Sandy arc 2945 Keewaywin platform 3050 to 2970 Sandborn platform 3) Muskrat Dam Lake Rottenfish arc	greenstones: E–W granitoids: EW to SE	2886(?)	N: <2730 >2699 S: ~2806	D.W. Davis, ROM, personal communication
southern subunit	N: Sandy Lake shear zone S: Unnamed shear zone at Pickle Lake E: North Caribou Lake–Totogan Lake shear zone W: Bear Head fault zone	1) North Caribou Lake Agutua Arm arc 2990 Keeyask platform 2980 detrital North Rim oceanic 2932 South Rim oceanic (?) 2) North Spirit Lake Makataiamik arc 3023 Disrupted arc 2900 to 2950 Nemakwis platform 2986 3) Muskrat Dam Lake Nekence platform 2958 to 2967 4) Horseshoe Lake Central Horseshoe arc ~2.9 Ga Horseshoe platform ~2.9 Ga Northern Horseshoe platform (?) Southern Horseshoe platform (?) Wapamisk arc ~2.9 Ga 5) Upper Windigo platform ~2.9 Ga 6) Favourable Lake Setting Net arc 2926	greenstones: NW granitoids: NW	2871		Corfu and Wallace 1986 DeKemp 1987 Corfu and Ayres 1990
Berens River subunit	N: Bear Head fault zone E: Narrow Lake Shear Zone S: Pipestone Bay–St. Paul Bay Deformation Zone	1) Red Lake 2989 to 2894 Ma oceanic 2) Confederation Lake (reference 1,2) 2960 Ma carbonate platform 3) McInnes Lake Western McInnes arc volcanism 2.9 Ga(?) Power oceanic volcanism 2.8 Ga(?) Southern McInnes arc volcanism 2.8 Ga(?)	1) east-trending complex folds 2) north-trending, east-younging homocline 3) north-trending, east-younging, west-younging	2870 to 2840 (reference 3)		1) Andrews et al. 1986 2) Corfu and Andrews 1987 3) Noble et al. 1989

Table 5.15. Terranes of the northwestern Superior Province.

Terrane	Boundaries	Greenstone Belts, Assemblage Types, Ages (Ma)	Structural Trends	Cratonization Age (Ma)	Docking Age (Ma)	References
Island Lake Terrane	N: unnamed shear zone South of Goose Lake to Playgreen Lake S: shear zone marked by ultramafic intrusions along central Island Lake	1) Stevenson Lake–Knight Lake–Hayes River Group arc >2895 2) Island Lake–Hayes River Group arc >2886	greenstones: EW granitoids: EW	2886 (?)	N: 2719 S: 2730 to 2699	Turek et al. 1986 Gilbert 1985
		3) tonalite gneiss with >2889 Ma amphibolite inclusions 4) microcline megacrystic intrusions 2720 (?) 5) late pyroxene monzodiorite 2705	greenstones: E W granitoids: E W			Turek et al. 1986 Corkery and Lenton 1989
Munro Lake Terrane	N: unnamed shear zone south end of Gods Lake S: unnamed shear zone south of Goose Lake to Playgreen Lake	1) Munro Lake basalts, minor rhyolites, cordierite-bearing wacke—non arc(?) 2) Goose Lake basalts, minor rhyolite, mafic + calcic metasediments 3) Cross Lake–Pipestone Group basalts, minor argillites, 2760—non arc (?) 4) Ponask-Sachigo Sachigo assemblage quartz-rich sediments, tholeiites, komatiites, no rhyolites—non arc 5) Koni Lake basalts, quartz-rich sediments—non arc	greenstones: SE to E granitoids: SE to E	2740 (?) (cut by Whisky Jack complex)	N: 2707 S: 2707 to 2720	Corkery and Lenton 1989 MEM 1987
Oxford Lake-Stull Lake terrane	N: boundary with Trans–Hudson Orogen indeterminate—north of Oxford Lake may be more Archean terranes S: Stull Lake–Wunnummin Lake fault zone	1) Oxford Lake–Knee Lake–Hayes River Group several mafic-felsic cycles, becomes calc-alkalic upward. 2830 Ma arc unconformably overlain by Oxford Lake Group 2707 Ma pull apart basin 2) Stull Lake Stull Lake arc Rapson pull apart (?) Richardson Arm pull apart (?) 3) Big Trout Lake arc within major layered intrusion	greenstones: EW to SE	2783 to 2730 (?)	N: (?) S: 2707	MEM 1987

Table 5.15. Terranes of the northwestern Superior Province.

Terrane	Boundaries	Greenstone Belts, Assemblage Types, Ages (Ma)	Structural Trends	Cratonization Age (Ma)	Docking Age (Ma)	References
Pickle terrane	1) Pickle Lake area N: Tectonic S: unknown	1) island arc 2892 Ma	ENE	2850 (?)	N: 2900 to 2850 Ma S: 2840 to 2806 Ma	Stott, this volume
	2) Red Lake area N: intrusive S: shear zone	2) island arc 2894 to 2892 Ma	NNW	—		
Woman assemblage	N: Windigo Lake–Horseshoe Lake shear zone S: Woman Lake Shear Zone shear zone in Lake St. Joseph area	1) Woman Lake 2830 Ma arc 2) Meen Lake 2840 to 2825 Ma arc 3) Pickle Lake arc 2890 Ma 4) North Spirit Lake, Makataiamik Lake assemblage ~2.8 Ga	greenstones: E–W trend, minor N–S granitoids: EW	2750 (?)	N: 2806 S: 2750	Corfu and Stott 1989 Stott and Wallace 1984
Muskrat Dam terrane	N: unnamed shear zone through Munekun Lake S: North Caribou Lake–Totogan Lake shear zone	1) Muskrat Dam Lake—Muskrat Dam assemblage 2734 Ma arc	E–W	<2730 Ma ~2710 Ma	N: 2730 to 2703 Ma S: 2690 to 2730 Ma	D.W. Davis, ROM, personal communication 1989

Sachigo Subprovince (as described by Card and Ciesielski 1986). There is some evidence suggesting that this terrane may in future be split into 2 subunits divided by the Sandy Lake shear zone. The following differences exist between the 2 potential subunits:

1. Cr and Ni abundances in quartz arenites differ, with elevated values south of the shear compared to those to the north
2. the slightly younger age of volcanism at Sandy Lake (2945 Ma) compared to the North Caribou arc sequence
3. the northwest trend of greenstone belts south of the shear compared with the westerly trend north of the shear zone

Episodic rejuvenation of the source area during sedimentation of the Keewaywin assemblage in the Sandy Lake greenstone belt indicates that the Sandy Lake shear zone is a possible tectonic boundary. Cortis (1988) described the Keewaywin assemblage as a basal coarsening-upward unit (shale and ironstone to arkosic sands) overlain by a fining-upward sequence (arenites to muds). The rejuvenation and the location along a major shear zone indicate that the assemblage may be a pull-apart basin. If this interpretation is valid, the Sandy Lake shear zone may be a terrane boundary. However, clear evidence for the age of the shear zone is lacking.

The Berens River Subprovince described by Card and Ciesielski (1986) is included within the North Caribou terrane because of the similarity between the 2.9 billion-year-old (Corfu and Wallace 1986) greenstones at Red Lake and those of the area to the west. The units at Red Lake have been termed carbonate-type platforms by Thurston and Chivers (1990). The greenstones of the McInnes Lake greenstone belt include quartz arenite-komatiite units similar to greenstone belts to the west (Cortis et al. 1988). Metamorphic data (Stone 1990) summarized above indicate that the Berens River Subprovince described by Card and Ciesielski (1986) represents the mid-crustal level of a granite-greenstone terrane. The Berens River Subprovince, as formerly designated, is separated from the North Caribou terrane by the Bear Head fault zone, but geochronological data (Corfu and Ayres 1984) suggest that movement along the fault is late, relative to greenstone ages (*see* Table 5.15) and the age of cratonization of the North Caribou terrane. The Bear Head fault zone therefore represents late adjustment to events possibly in southern Superior Province.

If the Berens River Subprovince is included within the North Caribou terrane, an explanation of the origin of the 2.8 and 2.7 billion-year-old greenstones within the Favourable Lake greenstone belt (Corfu and Ayres, in press) and the North Spirit Lake greenstone belt (Corfu and Wood 1986) is required. The alternatives are:

1. tectonic transport from the arc sequences of similar age within the Uchi Subprovince to the south (cf. Stott and Corfu, this volume)
2. The 2.8 and 2.7 billion-year-old sequences are along a suture represented by the Bear Head fault zone. This is a viable alternative, but it requires the Bear Head fault zone to be an early structure and does not explain the juxtaposition of 2.89 and 2.83 billion-year-old

sequences in the Pickle Lake area (Stott and Corfu, this volume).

3. The 2.7 and possibly the 2.8 billion-year-old sequences in the North Spirit Lake and Favourable Lake greenstone belts may represent ensialic greenstone belts. This is supported by the abundant quartz-arenite pebbles within the metasedimentary rocks of the Hewitt Lake Assemblage (*see* Table 5.4). The 2.8 billion-year-old inherited component in zircons of the 2.7 billion-year-old (Corfu and Wood 1986) Hewitt assemblage also suggests an ensialic origin.

The North Caribou terrane is bounded on the south by rocks assigned to the Uchi Subprovince by Card and Ciesielski (1986). The units forming the northern fringe of the Uchi Subprovince have been assigned to the Pickle terrane and the Woman assemblage by Stott (this volume). These units provide constraints on the timing of events that took place within the region examined in this chapter; therefore, they are briefly described later in this section.

Island Lake Terrane

The terrane is composed of arc greenstones extending from Knight Lake and Island Lake in Manitoba, through to Lingman Lake and the Pierce assemblage of the Pierce–Ponask–Sachigo greenstone belt in Ontario. The Ontario occurrences are correlated on the basis of rock types and structural trends with those at Island Lake in Manitoba. The arc greenstones of the Island Lake area have been termed the Hayes River Group and dated at more than 2886 Ma (Turek et al. 1986). The Hayes River Group is unconformably overlain by the laterally extensive 2768 to 2730 million-year-old Island Lake Series alluvial-fluvial sequence (Gilbert 1985). The Island Lake Terrane is distinguished from the North Caribou terrane to the south by the following:

1. the lack of quartz-rich epicratonic platform sequences
2. the slightly lower aeromagnetic intensity of granitic units
3. the slightly younger age of the greenstones and of post-tectonic granitic rocks

The character of the Island Lake Group is subject to variable interpretations. The group consists of conglomerates containing clasts which measure up to 3 m overlying a quartz wacke regolith. The map pattern, rock types and fault-bounded nature of the unit, coupled with the great lateral continuity of the conglomerates, suggest that it represents infilling of a continental rift. The group's position near the shear zone forming the south boundary of the Molson Domain (Manitoba Minerals Division 1987) and the presence of alluvial or mass flow conglomerates may indicate a possible late unconformable or pull-apart basin origin for the sequence. This relationship also indicates that the shear zone is a relatively late structure.

That part of the Island Lake Terrane which has been termed the Molson Domain in publications of the Manitoba Mines Branch (MMD 1987) consists of approximately 2.8 billion-year-old gneissic units and more than 2.84 billion-year-old supracrustal remnants of the Teacher Group (Corkery and Lenton 1989). The region is cut by approximately 2.7 billion-year-old potassic-feldspar megacrystic granitic plutons (Corkery and Lenton 1989). Granulite-facies gneisses occur on the south side of this part of the Island Lake Terrane within Manitoba (W. Weber, Manitoba Energy and Mines, personal communication, 1989) suggesting large-scale south-side-up movement. Alternative interpretations of the northern part of the Island Lake Terrane are:

1. a magmatic arc completely independent of the Island Lake Terrane to the south
2. a magmatic arc produced by 2.7 billion-year-old (MEM 1987) swamping of 2.8 billion-year-old Island Lake Terrane rocks.

The latter option is preferred. Note the similarity between this interpretation and the interpretation of the Berens River Subprovince as a mid-crustal level of the North Caribou terrane (cf. Stott and Corfu, this volume).

Muskrat Dam Terrane

This terrane contains complexly folded 2734 million-year-old (D.W. Davis, Royal Ontario Museum, personal communication, 1989) arc greenstones. It is in tectonic contact with the North Caribou terrane platformal sequences to the south, and with oceanic volcanic sequences of the Munro Lake Terrane to the north. Given these relationships, 2 interpretations are possible:

1. The magnetite-rich character of the granitic rocks east of Muskrat Dam Lake (similar to the Molson Lake Terrane granitoids) suggests a possible correlation with the Island Lake Series to the west.
2. The terrane may represent a suture between the 2760 million-year-old Munro Lake Terrane and the Island Lake Terrane. This 2.73 billion-year-old folded assemblage provides a maximum age for all accretion and folding events north of this terrane.

The suture hypothesis is preferred.

Munro Lake Terrane

This granite-greenstone terrane lies mainly within Manitoba, but a narrow sliver, including the Sachigo assemblage of the Pierce–Ponask–Sachigo greenstone belt and the Kino Lake belt (Hudec 1964), occurs within Ontario. The terrane differs from the other volcanic dominated greenstones in the following respects:

1. Metasedimentary rocks include calc-silicate units, euxinic units, and magnesium-rich cordierite-bearing wackes (MEM 1987).
2. Rhyolites are scarce.
3. Quartzose wackes and impure quartzites (Hudec 1964) form interbeds in the metavolcanic units.
4. Iron formation interbeds are common.

The volcanic sequences of the Munro Lake Terrane are similar to the mafic plain association described by Thurston and

Chivers (1990). The metavolcanic units are correlated with the 2760 million-year-old Pipestone Group (Corkery and Lenton 1989) at Cross Lake in Manitoba, based on structural trends and similar rock types. A minimum age for the shear zone boundary is provided by the 2713 million-year-old pluton at Walker Lake, which stitches the southern boundary of the terrane.

The terrane includes cross-bedded arkose, wacke, and related granodiorite-derived conglomerate (e.g., north of Goose Lake in Manitoba) similar in depositional environment to the 2706 million-year-old Oxford Lake Group. This indicates that the age of movement on the north boundary of the terrane is perhaps constrained to about 2706 Ma.

In a regional sense, the terrane is a suture between the reactivated 2.8 billion-year-old cratonic Molson Lake Terrane to the south and the 2.8 billion-year-old Oxford Lake–Stull Lake terrane to the north. The 2.7 billion-year-old (MEM 1987) potassic-feldspar megacrystic granitoids of the Molson Domain are of an appropriate age to represent back-arc plutonic activity during the late stages of south-directed 2.7 billion-year-old subduction related to the metavolcanic units of the Munro Lake Terrane.

Oxford Lake–Stull Lake Terrane

This is the northernmost terrane in Ontario, and is better exposed and more extensively mapped in Manitoba. The terrane is dominated by the Oxford Lake–Knee Lake greenstone belt and eastern extensions of the belt including all greenstones between Stull Lake and Big Trout Lake in Ontario. The terrane is dominated by the Hayes River Group, a 2.83 billion-year-old arc sequence unconformably overlain by the 2706 million-year-old Oxford Lake alluvial-fluvial pull-apart basin assemblage (MEM 1987). In Manitoba, Hubregtse (1978) has observed as many as 6 mafic to felsic volcanic cycles at southern Knee Lake. The entire terrane is constrained by a single age on each of the major groups as listed above and in Table 5.15. Given the similarity of rock types and structure to the Uchi Subprovince (*see* Stott and Corfu, this volume), characterization of the terrane as a 2.8 billion-year-old arc with a 2.7 billion-year-old pull-apart sequence at the southern margin is speculative. The Hayes River Group is intruded by tonalitic and granitic plutons of the Bayly Lake Complex dated at 2883, 2783 and 2730 Ma (MEM 1987), suggesting much complexity remains to be unravelled.

Pickle Terrane

The Pickle terrane defined by Stott (this volume) forms the southern margin of the North Caribou Terrane near Pickle Lake and at the east end of the Red Lake greenstone belt. The Pickle terrane consists of the Pickle Crow (2892 Ma) assemblage in the vicinity of Pickle Lake, and the Bruce Channel assemblage (*circa* 2894 Ma) of the Red Lake greenstone belt. Both assemblages are of similar age and are interpreted as arc volcanic assemblages. They are considered to be an arc terrane accreted onto the south margin of the North Caribou terrane during the period between 2890 and 2840 Ma.

Woman assemblage

The North Caribou terrane is bordered on the south by the Woman assemblage from the Confederation Lake area in the west, to the Hudson Bay Basin in the east (Figure 5.23; Stott, this volume). Stott (this volume), argues that the Woman assemblage is an autochthonous or parautochthonous continental arc-marginal basin complex on the south margin of a composite North Caribou–Pickle terrane. The Woman assemblage defines the southern limit of the relatively small and variable depositional settings of the terranes of the northwestern Superior Province which are described above.

The Woman assemblage is a possible source for the greenstone remnants at Favourable Lake (Corfu and Ayres, in press) and North Spirit Lake (Corfu and Wood 1986). The main evidence for the continental arc interpretation is the coeval 2838 million-year-old tonalite phase of the Trout Lake batholith, which intruded the base of the Balmer assemblage, suggesting the separation of a continental fragment and development of the Woman assemblage on rifted continental crust.

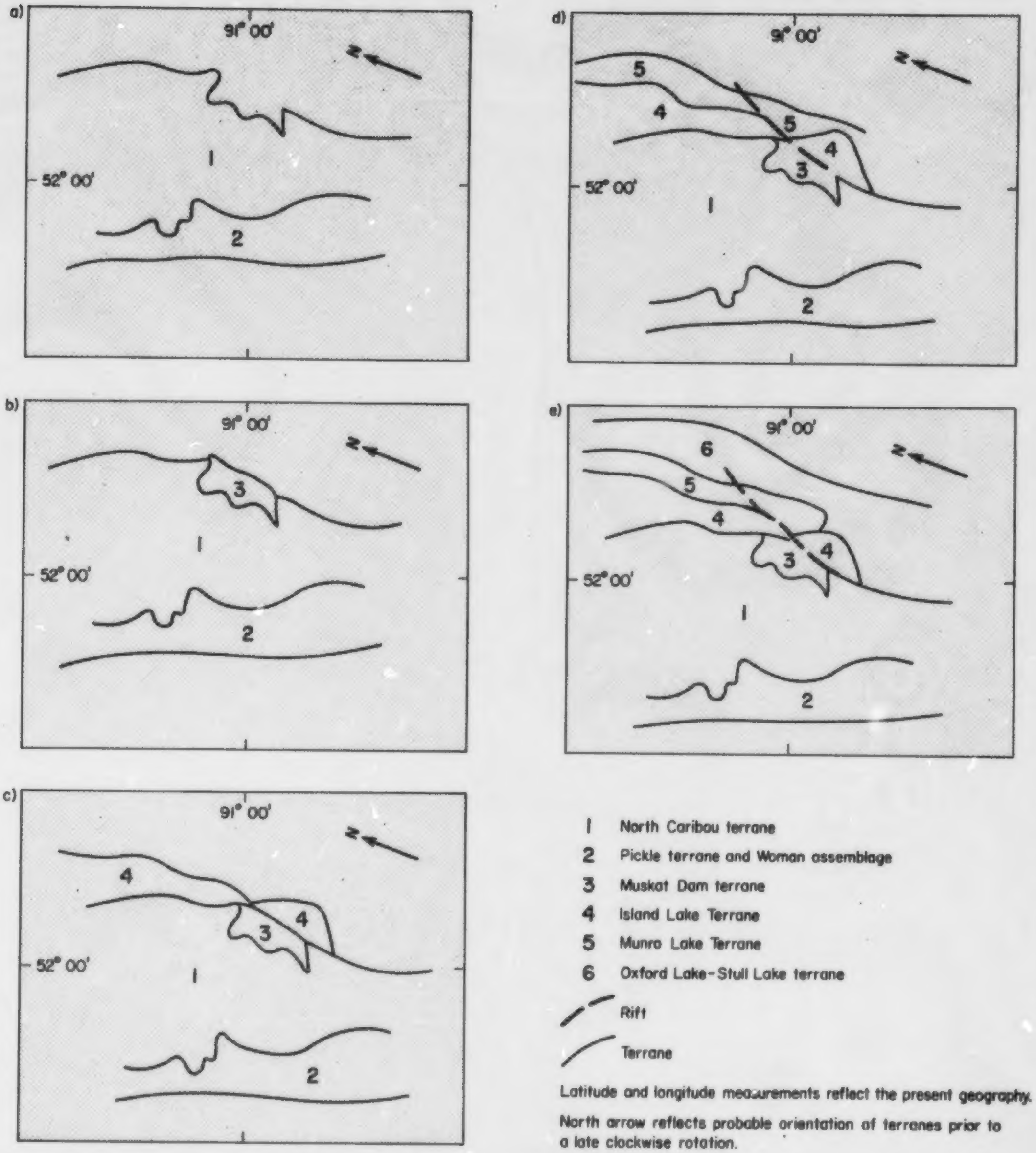
Alternatively, the Woman assemblage may be a conventional island arc accreted onto the south margin of the North Caribou terrane. If this is the case, the 2806 million-year-old (Noble et al. 1989) phase of the Trout Lake pluton may serve as a stitching pluton, constraining the age of juxtaposition of the North Caribou terrane and the Woman assemblage to the interval 2840 to 2806 Ma (Table 5.16).

TERRANE INTERACTION

The foregoing section has argued that each terrane is a distinctive entity in terms of lithostratigraphic assemblages, age of volcanism and cratonization. Where geological relations are well established, the terranes are bounded by regional-scale shear zones. Our method of determining the age of juxtaposition of these terranes is constrained by a variety of approaches. Criteria used are more general than those used in the pioneering study of Corfu and Stott (1986) because of the early stage of geochronological and geological investigations in northwestern Superior Province. Criteria used to constrain the age of shear zone development, enumerated earlier, include the ages of pull-apart basin and molasse basin assemblages, the ages of cratonization of terranes versus the ages of volcanism in adjacent terranes, and certain geometric constraints. The ages of terrane juxtaposition and interaction are discussed in approximate order of decreasing age. A more detailed rationale is listed in Table 5.16. The process of terrane interaction is shown in Figure 5.24. The details are portrayed on Figure 5.25, a time-space diagram.

The earliest event may be the interaction between the northern and southern subunits of the North Caribou terrane occurring between 2970 and 2950 Ma (D.W. Davis, Royal Ontario Museum, personal communication, 1991).

The age of this event is based (Figure 5.26) on the limited constraints on the age of the Keewaywin assemblage pull-apart basin. The Woman assemblage docked against the North Caribou terrane at about 2806 Ma. The 2734 mil-


a)
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1 North Caribou terrane
2 Pickle terrane and Woman assemblage
3 Muskat Dam terrane
4 Island Lake Terrane
5 Munro Lake Terrane
6 Oxford Lake-Stull Lake terrane
Rift
Terrane

Latitude and longitude measurements reflect the present geography.

North arrow reflects probable orientation of terranes prior to a late clockwise rotation.

Figure 5.24. Cartoon showing the sequence and age of terrane interaction events for the northwestern Superior Province. **a)** 2806 Ma—Accretion of Woman assemblage against North Caribou terrane. **b)** 2734 to 2720 Ma—Docking of Muskrat Dam terrane against North Caribou terrane. **c)** 2734 to 2720 Ma—Accretion of Island Lake Terrane. **d)** 2730 to 2719 Ma—Docking of Munro Lake Terrane against Island Lake Terrane. Rifting occurs within the 2 terranes. **e)** 2709 Ma—Accretion of Oxford Lake–Stull Lake terrane against Munro Lake Terrane. Terranes are shown in greater detail in Figure 5.23.

lion-year-old Muskrat Dam Lake greenstones were juxtaposed between 2734 and 2720 Ma. This was followed by the accretion of the Island Lake and Munro Lake terranes on the north, during the interval 2734 to 2720 Ma. The Oxford Lake–Stull Lake terrane was juxtaposed with the terranes to the south at 2707 Ma.

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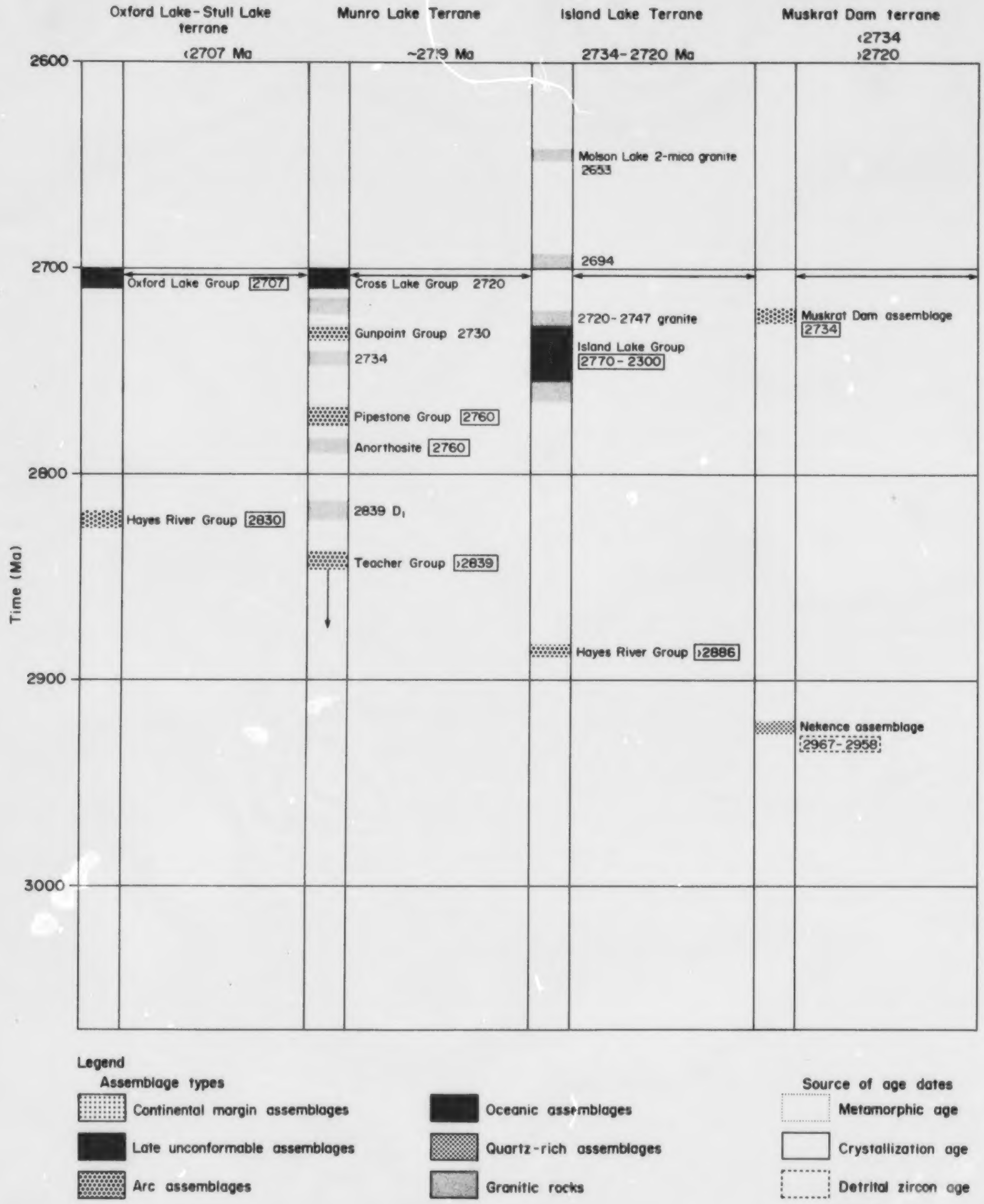
Oxford Lake-Stull Lake terrane
‹2707 Ma
Munro Lake Terrane
~2719 Ma
Island Lake Terrane
2734-2720 Ma
Muskrat Dam terrane
‹2734
›2720
2600
2700
2800
2900
3000
Time (Ma)
Oxford Lake Group 2707
Hayes River Group 2830
Cross Lake Group 2720
Gunpoint Group 2730
2734
Pipestone Group 2760
Anorthosite 2760
2839 D₁
Teacher Group ›2839
Molson Lake 2-mica granite
2653
2694
2720-2747 granite
Island Lake Group
2770-2300
Hayes River Group ›2886
Muskrat Dam assemblage
2734
Nekence assemblage
2967-2958
Legend
Assemblage types
Continental margin assemblages
Late unconformable assemblages
Arc assemblages
Oceanic assemblages
Quartz-rich assemblages
Granitic rocks
Source of age dates
Metamorphic age
Crystallization age
Detrital zircon age

Figure 5.25. Generalized time-space diagram for the northwestern Superior Province. Terranes largely outside Ontario are not subdivided into assemblages. The horizontal arrows indicate the approximate age of terrane juxtaposition

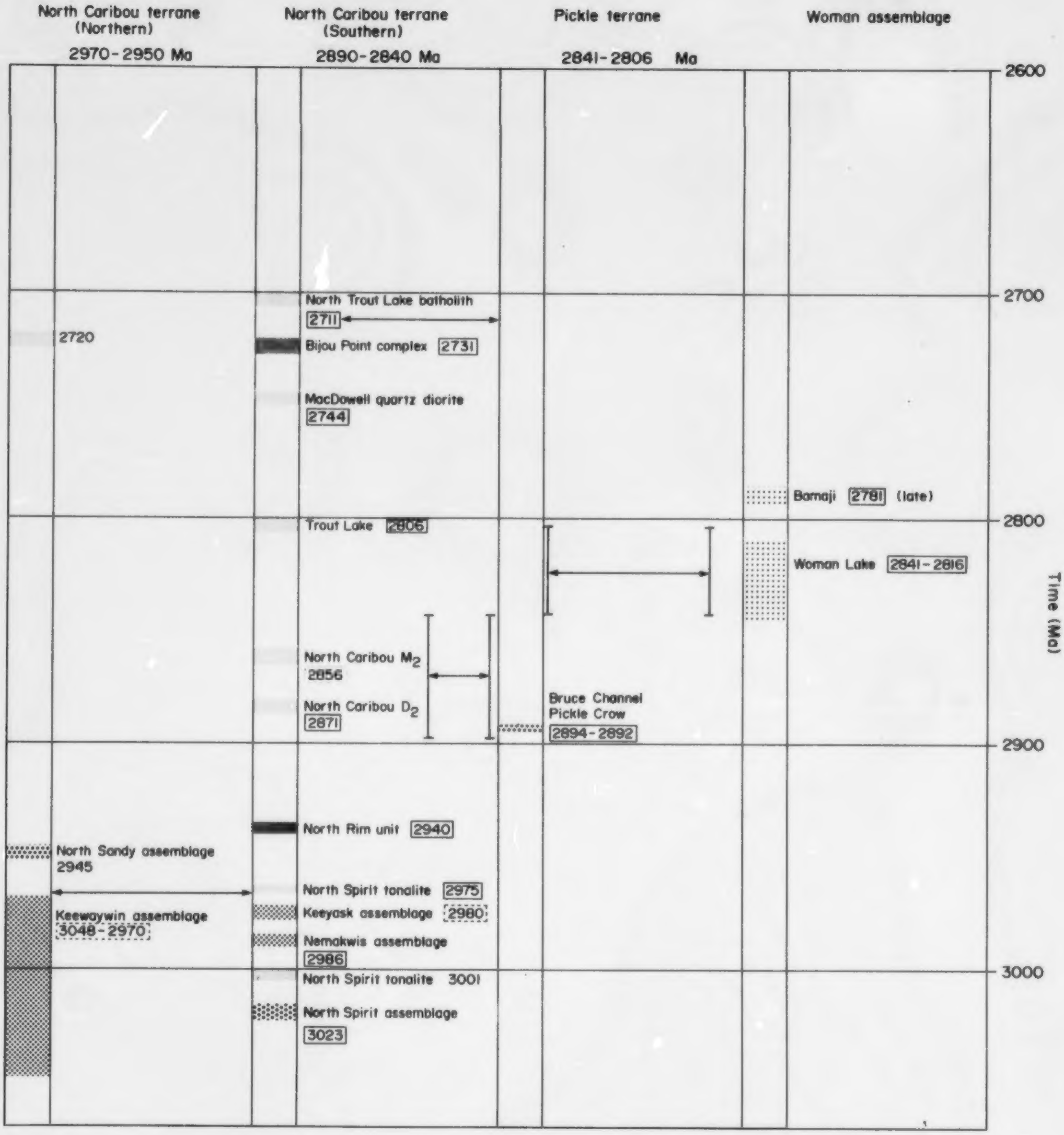

Figure 5.25. Generalized time-space diagram for the northwestern Superior Province. Terranes largely outside Ontario are not subdivided into assemblages. The horizontal arrows indicate the approximate age of terrane juxtaposition

Ontario would not have been possible without the additional data acquired through 2 seasons of reconnaissance evaluation (Thurston et al. 1987; Cortis et al. 1988) and the rethinking of the role of shear zones (Osmani and Stott 1988) which benefitted from the wise counsel of G.M. Stott. The following members of the field crew involved in the re-evaluation helped to cover an enormous area under difficult conditions: K.M. Chivers, A.L. Cortis, D. Aiken and T. Hoey.

The study benefitted enormously from the unpublished geochronological results furnished by D.W. Davis of the Jack Satterly Geochronology Laboratory of the royal Ontario Museum. We acknowledge the benefits we obtained from several discussions of geology on the other side of the "line" with W. Weber and especially M.T. Corkery of the Geological Services Branch, Manitoba Department of Mines and Energy. The native communities that provided access to their lands helped our understanding of the area greatly.

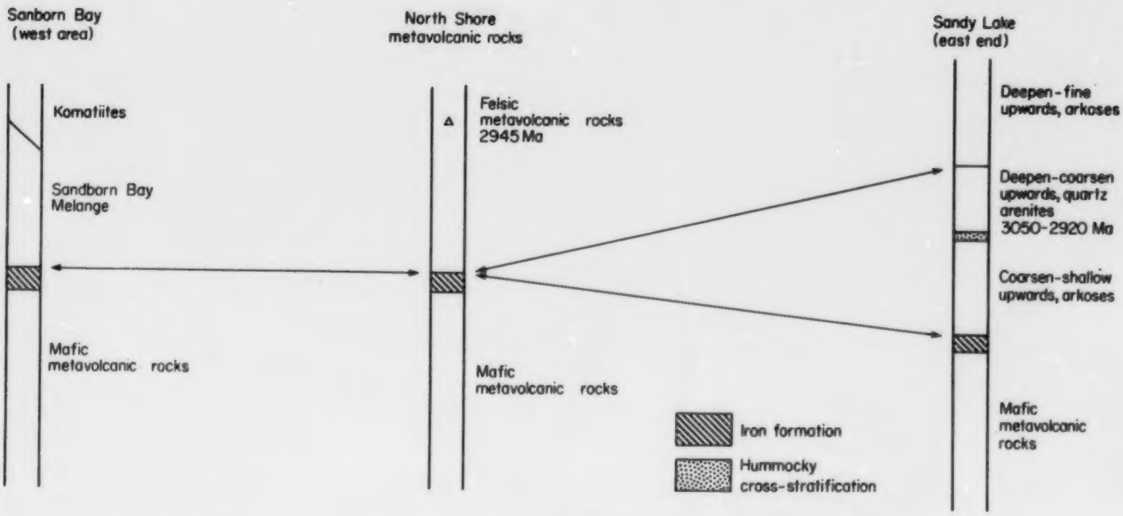
Sanborn Bay (west area)
Komatiites
Sandborn Bay Melange
Mafic metavolcanic rocks
North Shore metavolcanic rocks
Felsic metavolcanic rocks 2945 Ma
Mafic metavolcanic rocks
Iron formation
Hummocky cross-stratification
Sandy Lake (east end)
Deepen-fine upwards, arkoses
Deepen-coarsen upwards, quartz arenites 3050-2920 Ma
Coarsen-shallow upwards, arkoses
Mafic metavolcanic rocks

Figure 5.26. Schematic columnar sections of the Sandy Lake greenstone belt showing tentative regional correlation of assemblages in the Sandy Lake greenstone belt based on mapping carried out by Thurston et al. (1987), Satterly (1939) and Cortis (1988), and geochronologic data provided by D.W. Davis (Royal Ontario Museum, written communication, 1990).

Table 5.16. Constraints on age of interterrane shear zone development.

Rationale	Age Constraints (Ma)	References
1) **North Caribou terrane** northern versus southern subunits juxtaposition produced the Keewaywin pull-apart basin which is >2945 Ma (Figure 5.3), but <2970 to 3048 Ma based on detrital zircons in the unit	max. 2970 min. 2950	D.W. Davis, ROM personal communication 1989
2) **North Caribou terrane versus Woman assemblage** Boundary between 2.9 Ga Balmer assemblage at Confederation Lake and 2.8 Ga Woman assemblage is a 2806 Ma stitching pluton	max. 2840 min. 2806	Nunes and Thurston 1980; Noble et al. 1989
3) **North Caribou terrane versus Muskrat Dam terrane** Accretion of the Muskrat Dam arc postdates the 2734 Ma age of the arc rhyolites and is probably prior to the shear–zone-related 2720 Ma units of the along the Island Lake–Munro Lake Terrane boundary	max. 2734 min. 2720	D.W. Davis, ROM personal communication 1989 Turek et al. 1986
4) **Muskrat Dam terrane versus Island Lake Terrane** Late folding of the Muskrat Dam assemblage of the Muskrat Dam Lake greenstone belt is younger than the 2734 Ma age of the rhyolites and prior to the age of posttectonic 2699 Ma dikes at Island Lake	max. 2734 min. 2729	D.W. Davis, ROM personal communication 1989 Turek et al. 1986
5) **Island Lake Terrane versus Munro Lake Terrane** Munro Lake Terrane greenstones are correlative with the 2760 Ma Pipestone Group and the 2730 Ma Gunpoint Group; shear zones cutting these sequences are ~2719 Ma.	max. 2730 min. 2719	Corkery and Lenton 1989
6) **Munro Lake Terrane versus Oxford Lake–Stull Lake terrane** Constrained by the 2707 Ma age of the Oxford Lake Group pull-apart basin sequence and the 2692 Ma age of a cross-cutting granitic unit	max. 2707 min. 2692	MEM 1987
7) **Pickle terrane versus North Caribou terrane** Constrained by 2894 to 2892 Ma age of Pickle Crow assemblage and the 2840 Ma age of the oldest Woman assemblage supracrustal rocks	max. 2892 min. 2841	Stott, this volume

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Chapter 6

Uchi Subprovince

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Abstract

The Uchi Subprovince contains a linear, belt-like collage of volcanic and sedimentary assemblages that represent discrete magmatic and erosional pulses during approximately 280 million years of Archean history. These supracrustal rocks, underlain by synvolcanic plutons, were invaded by younger felsic plutons and were shouldered aside by buoyant batholithic complexes during several orogenic periods, the most prominently preserved being the Kenoran Orogeny, which culminated in this part of the Superior Province about 2.7 Ga. Some clastic and chemical sedimentary sequences comprise the youngest units in the volcanic assemblages. Other sedimentary rocks form separate assemblages lying unconformably upon the volcanic units and formed mainly during the Kenoran Orogeny. Some volcanic assemblages are dominantly composed of tholeiitic basalt and komatiitic rocks, interpreted to have originated as oceanic mafic plain sequences, probably in a back-arc setting; most assemblages are composed of cycles or sequences comprising tholeiitic basalt platforms overlain by calc-alkalic andesite, dacite and rhyolite, interpreted to have originated in continental or oceanic arcs. The U-Pb isotopic ages of felsic volcanic rocks permit us to recognize that the present stacking of volcanic strata locally forms repetitions of individual volcanic cycles (e.g., the Confederation assemblage) or locally forms volcanic sequences stacked out of normal stratigraphic order, presumably by thrusting, as in the Confederation and St. Joseph assemblages.

Distinct orogenic periods are each interpreted to compose all or most of the following sequence of magmatic, sedimentary and structural events: a) intrusion of synvolcanic plutons into older volcanic sequences, in places stitching 2 assemblages and providing age constraints on the sequence of events; b) thrust stacking and tilting of volcanic and sedimentary strata within assemblages; c) erosive exhumation of deformed strata and plutons, with detritus collecting in synorogenic troughs; d) continued regional, horizontally directed shortening; and e) emplacement of late to posttectonic plutons. Only the final orogenic event (the Kenoran Orogeny) can be clearly shown to terminate with regional transcurrent displacement (transpression) across faults accompanied by crustal block rotation. Post–Archean faulting has occurred at least locally.

The collage of assemblages in the subprovince shows a general southward younging; one could postulate from the geological patterns and sequence of events that the assemblages of the Uchi Subprovince reflect episodic additions of crustal units to the margin of a growing continent. More complex age distributions and dramatic changes in stratigraphic orientations are interpreted to reflect allochthonous or parautochthonous, northward transport of some assemblages. Accordingly, this chapter offers a revised perspective of the Uchi Subprovince, which is viewed, not as a separate tectonic entity implied by its label as a subprovince, but as a leading edge of an evolving microcontinent that developed over 290 million years. This microcontinent also included the Berens River Subprovince and much of the southern Sachigo Subprovince.

A three-stage process of Archean continental growth is proposed. The northern parts of the Uchi Subprovince and the Red Lake greenstone belt, consisting of older volcanic assemblages, are interpreted as parts of a microcontinent comprising the North Caribou terrane which straddles the Uchi, Berens River and Sachigo subprovinces, and the Pickle terrane. These terranes accreted to form a composite terrane or microcontinent. The southern half of the Uchi Subprovince consists of episodic crustal additions to this microcontinent along its leading edge in the form of arc and marginal basin complexes and associated plutons. Hence, the North Caribou terrane accumulated additions to its northern and southern flanks to form the Uchi–Sachigo superterrane (a microcontinent), probably by successive accretion of suspect terranes (stage 1), and particularly in the Uchi Subprovince, by outward growth through episodic volcanism along an active, continental margin (stage 2) analogous to the Phanerozoic evolution of the Andes. The processes of continental growth were probably driven by mantle convective transport, comparable to processes along modern convergent plate margins. This growth along a convergent, subduction margin terminated when the microcontinent (the Uchi–Sachigo superterrane) and its fore-arc accretionary prism (the English River assemblage), collided with another microcontinent (converging from the "south") during the Kenoran Orogeny (stage 3). This collision marked one phase in a rapid progression of superterrane collisions (i.e., of subprovinces and sets of subprovinces) leading to the construction of a stable craton—the Superior Province.

INTRODUCTION

Gold exploration began to open up the Uchi Subprovince to the scrutiny of geologists and prospectors. Major discoveries of gold by far-ranging prospectors attracted the interest of others and stimulated several staking rushes during the 1920s and 1930s. The Ontario government sent geologists to investigate reports of volcanic rocks in this region and to conduct reconnaissance and detailed mapping of parts of this subprovince as it opened up. This process continues to this day since much of the northwestern part of the province, isolated from road access, has nonetheless received additional mapping to encourage continued exploration for gold and volcanogenic massive sulphide deposits.

Since the early years of discovery and exploration, we have come to recognize that the Uchi Subprovince of Card and Ciesielski (1986; Figure 6.1) is a long, slender volcanic and plutonic domain. It occupies a tectonically significant position along the southern flank of a cratonic block, which grew to encompass the Sachigo, Berens River and parts of the Uchi subprovinces, as they are defined by Card and Ciesielski (1986), and is composed of domains of rocks older than 2800 Ma. This block is distinguished from the block of subprovinces to the south, which is composed of eastward-trending terranes dominated by younger volcanism, erosion and sedimentation. In the Uchi Subprovince, the main interconnecting volcanic belts encompass a breadth of up to 80 km. The length of this subprovince is exposed in northwestern Ontario for 600 km and continues farther westward into Manitoba for another 90 km where it becomes overlain by Phanerozoic rocks at Lake Winnipeg. It is overlain to the east by the Phanerozoic rocks of the Hudson Platform, but may correlate with the La Grande River Subprovince, east of James Bay in northern Quebec. The apparent full extent, correlating Uchi–La Grande River subprovinces, from the trace of the Proterozoic Trans–Hudson Orogen in Manitoba to the Labrador Trough in Labrador, is approximately 2000 km. The length of such a belt-like system of arc volcanism and sedimentation, bordering a terrane of clastic sedimentary rocks, compares in scale with Phanerozoic volcanic arcs and bordering accretionary prisms (cf., Hamilton 1988).

The approach taken in this chapter is to summarize the stratigraphic characteristics of the tectonic assemblages (defined below) in each of the greenstone belts from west to east, and to review the structural characteristics, plutonic suites and the economic geology of the subprovince as a whole, using examples from different parts of the subprovince. The belts are named as shown on Figure 6.2 and where they merge into each other, an arbitrary subdivision for descriptive purposes is made in the vicinity of narrow supracrustal transitions between the belts. Later, a summary and analysis of the tectonic events is given, by integrating the geology of these assemblages and plutonic suites, the structural framework and available age determinations. Age determinations are taken from OGS Map 2576 of the Tectonic Assemblages of Ontario (*see* Map 2576, map case) and from charts (*see* Maps 2579 and 2580, map case). Most of the unpublished age determinations in the Uchi Subprovince are those of the second author, F. Corfu of the Jack Satterly Geochronology Laboratory, Royal Ontario Museum.

Since all supracrustal rocks, volcanic and sedimentary, plus the synvolcanic and pre- to syntectonic plutonic rocks are metamorphosed, the prefix **meta-** has been deleted to ease readability of the text. The late to posttectonic plutons that are essentially unmetamorphosed will be identified. Where reference is given to other subprovinces, notably the Berens River and Sachigo subprovinces, the names are used to refer to areas defined by Card and Ciesielski (1986) for ease of reference. It shall be proposed in this chapter and by Thurston, Osmani et al. (this volume) that terrane boundaries transect subprovince boundaries; for example, the 2.9 to 3.0 billion-year-old rocks of the western part of the Uchi Subprovince are considered part of the North Caribou terrane, which dominates the Sachigo Subprovince.

SUPRACRUSTAL ROCKS

Regional Stratigraphic Correlations—An Historical Perspective

It has long been assumed by some Archean geologists that greenstone belts are not separately evolved entities, but are remnants of previously more extensive supracrust that has undergone tectonism (Macgregor 1951). Yet it has more widely been proposed that granitic batholiths are the principal cause of the regional shapes and internal deformation of greenstone belts, either by forceful intrusion (Ramsay 1975) or by buoyant displacement of denser, volcanic belts (West 1980; Schwerdtner et al. 1979). In the past 10 years particularly, the results of several investigations in the Uchi Subprovince support the first assumption and challenge the second. We shall look at the evidence for this.

Part of the impetus for stratigraphic mapping arose from a widely accepted suspicion that both volcanic massive suphide and gold deposits were in part stratigraphically controlled. This has led to regional stratigraphic and geochemical correlation studies in different belts within Ontario such as Trowell et al. (1980) and Blackburn et al. (1985) in the western Wabigoon Subprovince and Wallace et al. (1986) in the western Uchi Subprovince. These studies benefitted greatly from U-Pb zircon age determinations, particularly of felsic to intermediate volcanic rocks. Absolute ages provided necessary constraints to stratigraphic correlations in areas of complexly interleaved strata. For example, in the Uchi Subprovince, the earliest attempt at regional correlation of volcanic strata between major greenstone belts was made by Thurston (1981). His efforts to trace volcanic cycles from the Birch–Uchi greenstone belt (*see* Figure 6.2) to the southern part of the Red Lake greenstone belt were substantially revised by Wallace et al. (1986) when more age determinations for rocks of the Red Lake greenstone belt became available (Corfu and Wallace 1986). Stott (Stott and Wallace 1984; Stott 1985, 1986) continued this program of regional stratigraphic correlation farther east in the Lake St. Joseph–Pickle Lake area (*see* Figure 6.2) and this has led to a general synthesis of the geology of the Uchi Subprovince as shown on Figure 6.2, and a synthesis of tectono-stratigraphic assemblages as shown in Figure 6.3. The latter synthesis is based on a compilation of geological mapping at all scales, numerous age determinations and an interpretation of aeromagnetic and electromagnetic maps. Figure 6.3 is an interpretive map, but sufficient confidence in the correlations exist to subdivide the supracrustal rocks into several tectonic assemblages (defined *in* Archean Geology of Ontario: Introduction).

The assemblage descriptions that follow are natural and useful subdivisions of the regional stratigraphy. This framework of assemblages is open to revision as more becomes

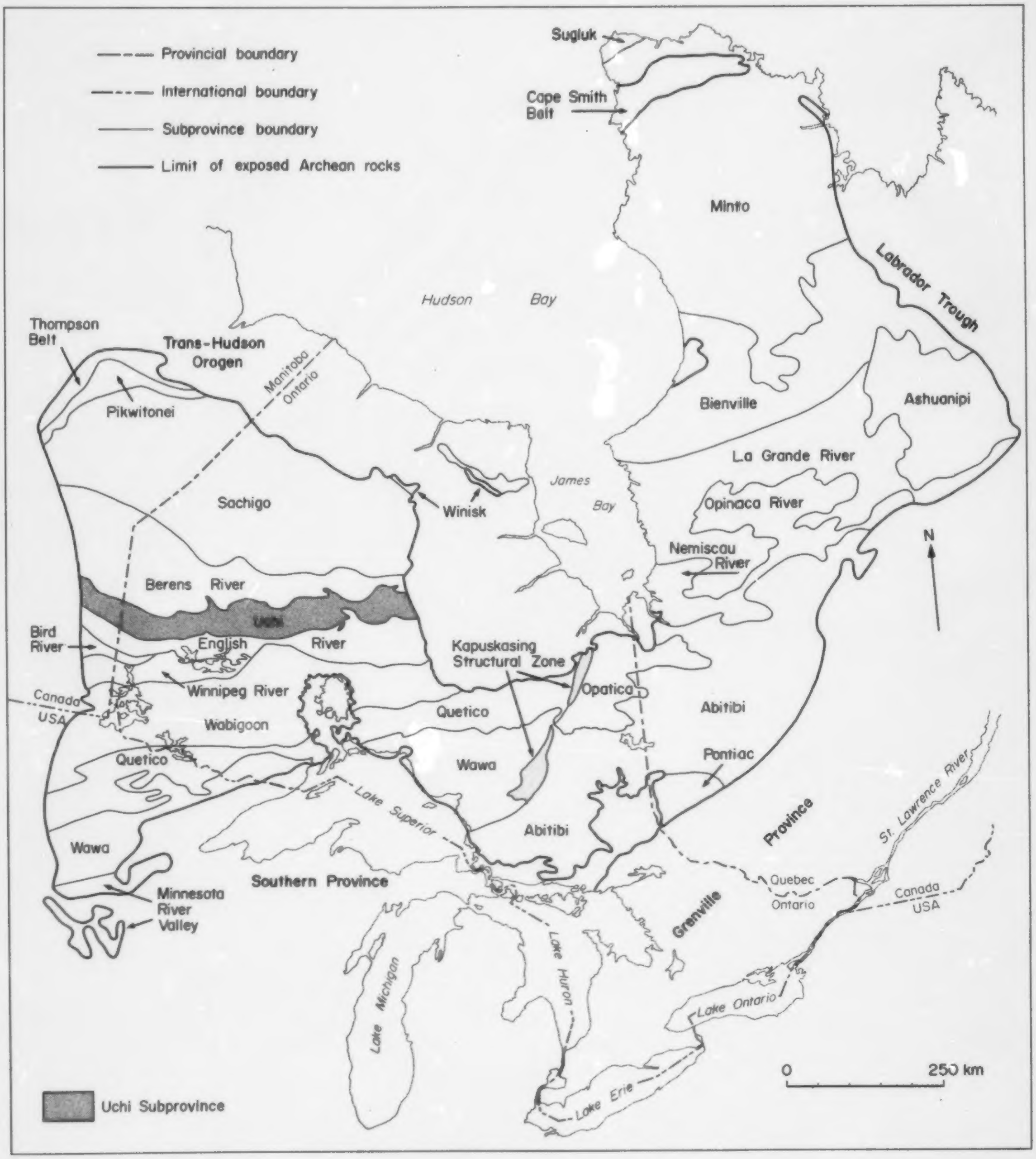

Provincial boundary
International boundary
Subprovince boundary
Limit of exposed Archean rocks
Sugluk
Cape Smith Belt
Minto
Labrador Trough
Hudson Bay
Thompson Belt
Trans-Hudson Orogen
Manitoba
Ontario
Pikwitonei
Bienville
Ashuanipi
La Grande River
James Bay
Sachigo
Winisk
Opinaca River
Nemiscau River
N
Berens River
Uchi
Bird River
English River
Kapuskasing Structural Zone
Opatica
Winnipeg River
Canada
USA
Wabigoon
Quetico
Abitibi
Quetico
Wawa
Pontiac
Lake Superior
Province
St. Lawrence River
Abitibi
Wawa
Southern Province
Quebec
Ontario
Canada
USA
Minnesota River Valley
Grenville
Lake Huron
Lake Michigan
Lake Ontario
0
250 km
Lake Erie
Uchi Subprovince

Figure 6.1. Location of the Uchi Subprovince in relation to other subprovinces in the Superior Province of Ontario.

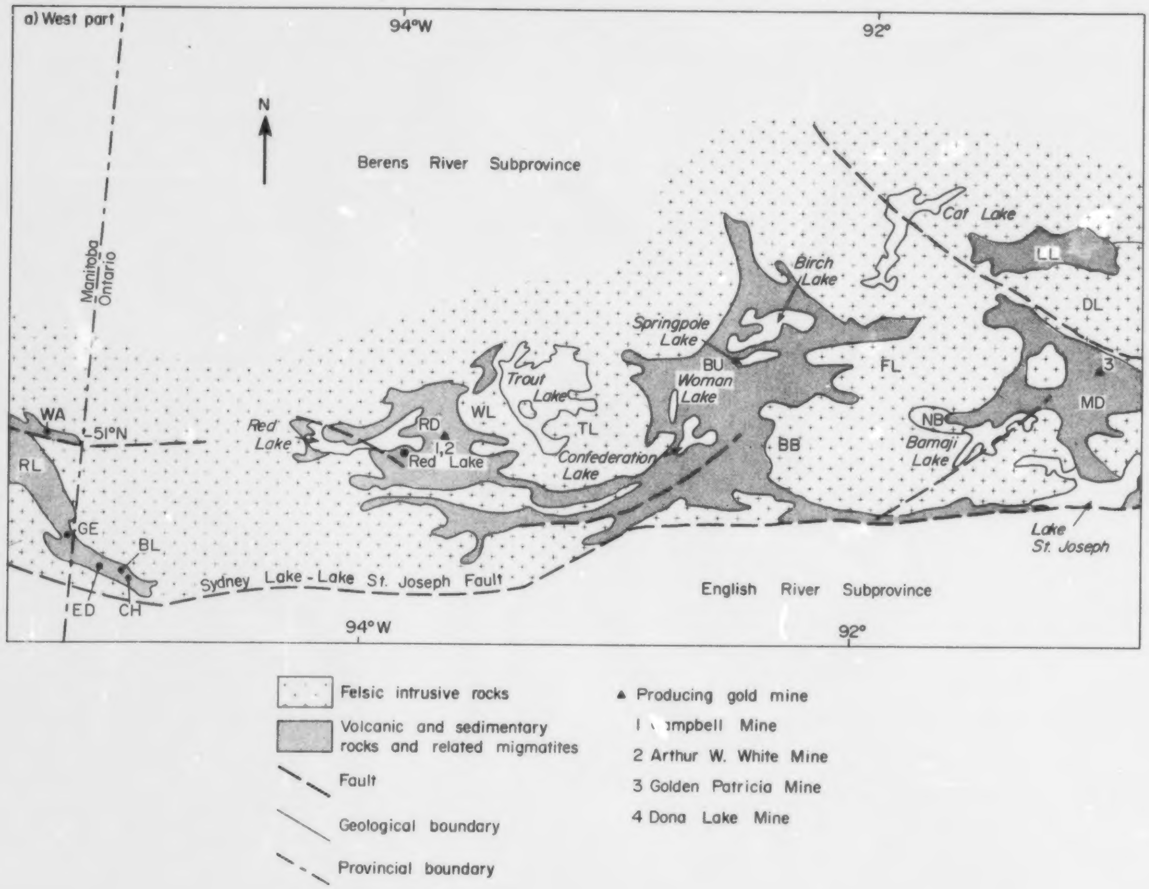

a) West part
94°W
92°
N
Berens River Subprovince
Manitoba
Ontario
Cat Lake
LL
Birch Lake
DL
Springpole Lake
BU
Woman Lake
FL
3
MD
Trout Lake
WL
RD
TL
NB
Bamaji Lake
WA
51°N
Red Lake
1,2
Red Lake
Confederation Lake
BB
RL
GE
BL
Lake St. Joseph
Sydney Lake - Lake St. Joseph Fault
English River Subprovince
ED
CH
94°W
92°
Felsic intrusive rocks
Volcanic and sedimentary rocks and related migmatites
Fault
Geological boundary
Provincial boundary
▲ Producing gold mine
1 Campbell Mine
2 Arthur W. White Mine
3 Golden Patricia Mine
4 Dona Lake Mine

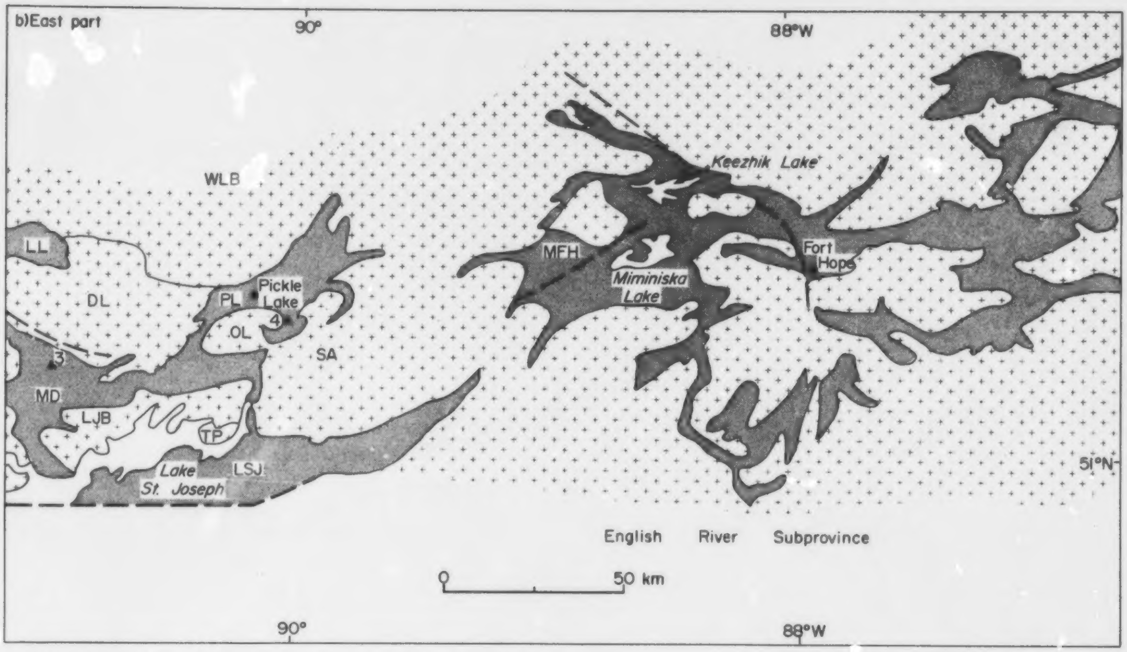
Figure 6.2. Generalized map of greenstone belts and granitic batholiths of the Uchi Subprovince, showing names of the belts, batholiths, and principal towns and lakes.

known about this region. For example, correlations of assemblages are tentative in the less well known, northern part of the Miminiska–Fort Hope greenstone belt (*see* Figure 6.2). In some areas, such as in the northern Pickle Lake greenstone belt, lack of adequate information inhibits either assigning such areas to existing assemblages or further subdivisions of the areas into assemblages; these areas are therefore left unnamed.

Tectonic Assemblages

The following sections provide a geological overview of the bedrock within the Uchi Subprovince. The descriptions and ensuing discussions are given under the headings of the 2 fundamental components of the subprovince—the greenstone belts and the batholiths external to these belts. Tectonic assemblages form the basic tectonostratigraphic subdivisions of the greenstone belts and are therefore the principal headings of stratigraphic descriptions.

Boundaries between adjacent assemblages are disconformable, faulted, or interfaces between adjacent volcanic cycles. But, the identification of these assemblages depends upon correlating stratigraphic units of volcanic cycles, or correlating marker beds in thick mafic plain sequences; and generally, only the upper felsic units of volcanic cycles or thin felsic units in mafic plain sequences have been dated. Where the position of the boundary between assemblages is not clear in areas where there is very little information and where an interpreted interface is inferred between assemblages, the less clearly defined boundaries are drawn parallel to stratification. In such cases, the assemblage boundary, based upon the concept of cyclical volcanism in an arc setting (cf., Thurston 1986), is tentatively placed between the upper felsic unit of a significantly older assemblage and the lower mafic unit of a younger assemblage. Boundaries between assemblages are typically not exposed in this region of extensive, albeit thin, glacial overburden and coniferous forests. Interpretations of assemblage contacts therefore depend upon several observations including: detailed descriptions of typically faulted but only locally exposed contacts; regional scale unconformities, typically best seen from the map patterns since locally exposed unconformities are rarely observed; contrasting trends of geophysical conductors; contrasting structural styles;



Abbreviations

BB	Bamaji-Blackstone batholith	BL	Bee Lake belt	BU	Birch-Uchi greenstone belt
CH	Chase Lake area	DL	Dobie Lake pluton	ED	Eden Lake area
FL	Fawcett Lake batholith	GE	Gem Lake area	LJB	Lake St. Joseph batholith
LL	Lang Lake greenstone belt	LSJ	Lake St. Joseph greenstone belt	MD	Meen-Dempster greenstone belt
NB	North Bamaji pluton	OL	Ochig Lake pluton	MFH	Miminiska-Fort Hope greenstone belt
PL	Pickle Lake greenstone belt	RD	Red Lake greenstone belt	RL	Rice Lake greenstone belt
SA	Seach-Achapi batholith	TL	Trout Lake batholith	TP	Twiname stock
WA	Wallace Lake area	WL	Walsh Lake pluton	WLB	Williams Lake batholith

Figure 6.2. Generalized map of greenstone belts and granitic batholiths of the Uchi Subprovince, showing names of the belts, batholiths, and principal towns and lakes.

contrasting stratigraphic younging directions; and contrasting lithological characteristics.

Assemblages have been used as informal subdivisions of greenstone belts, and are intended to assist us in testing regional-scale stratigraphic relationships and identifying paradoxical patterns. In this respect, assemblages serve a role as subdivisions of tectonically affected Archean stratigraphy; this is a role served more usefully and flexibly than by formations and groups, which tend to be assigned on a more local scale. This approach of dividing greenstone belts into assemblages is flexible because, by the definition applied in this volume (*see* Archean Geology of Ontario: Introduction) and the accompanying map of Tectonic Assemblages of Ontario, it permits us to identify assemblages as areas where greater precision and detail concerning the depositional environment are available, as in parts of the Abitibi greenstone belt. It also permits us to assign areas of more than one depositional environment to assemblages in areas of northwestern Ontario where at present further subdivision may not be possible to the same level of discrimination. This limited discrimination in many parts of the Uchi Subprovince is due in part to a limited geochemical data base. Assemblage characteristics on a regional scale are therefore largely based on lithological and structural features. From these features one can distinguish broad parts of the greenstone belts that bear characteristics similar to various depositional settings in modern crustal settings as described by Thurston and Chivers (1990). For example, a large part of the Red Lake greenstone belt is composed of subaqueous komatiitic and tholeiitic basalt flows, interrupted only by limited iron formation beds and distal facies tuff units; these areas resemble oceanic mafic plains that contrast markedly with other parts of the belt dominated by shallow water volcanic deposits. These latter areas are lithologically more complex and resemble modern island arc and continental arc settings. The metasedimentary rocks comprise a substantial component of the greenstone belts in the Uchi Subprovince. Contrasts in thickness and facies association reflect different environments of deposition. The metasedimentary sequences include:

1. chemical sediments marking breaks in volcanic activity and occurring at the top of basaltic flows in broad

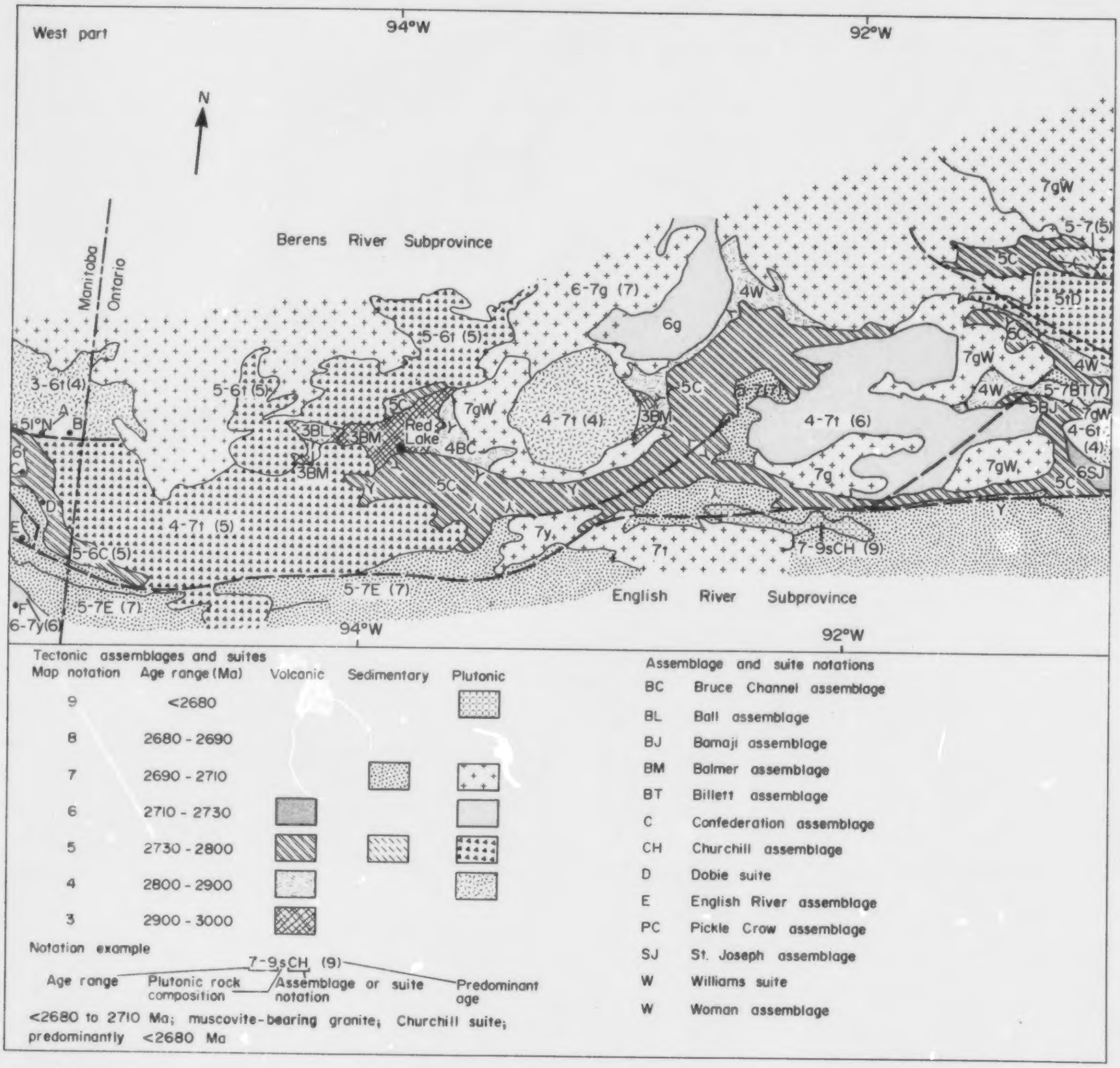

West part
94°W
92°W
N
Manitoba
Ontario
Berens River Subprovince
English River Subprovince
Red Lake
51°N
6-7g (7)
6g
4W
5C
5tD
7gW
5-7(5)
5-6t (5)
5-6t (5)
3-6t (4)
3BL
3BM
3BM
4BC
7gW
4-7t (4)
3BM
6-7t (7)
4-7t (6)
4W
5-7BT (7)
5BJ
7gW
4-6t (4)
6SJ
7gW
5C
7g
Y
4-7t (5)
5-6C (5)
7y
7t
7-9sCH (9)
5-7E (7)
5-7E (7)
6-7y (6)
A
B
C
D
E
F
Tectonic assemblages and suites
Map notation
Age range (Ma)
Volcanic
Sedimentary
Plutonic
9 <2680
8 2680 - 2690
7 2690 - 2710
6 2710 - 2730
5 2730 - 2800
4 2800 - 2900
3 2900 - 3000
Notation example
7-9sCH (9)
Age range
Plutonic rock composition
Assemblage or suite notation
Predominant age
<2680 to 2710 Ma; muscovite-bearing granite; Churchill suite; predominantly <2680 Ma
Assemblage and suite notations
BC Bruce Channel assemblage
BL Ball assemblage
BJ Bamaji assemblage
BM Balmer assemblage
BT Billett assemblage
C Confederation assemblage
CH Churchill assemblage
D Dobie suite
E English River assemblage
PC Pickle Crow assemblage
SJ St. Joseph assemblage
W Williams suite
W Woman assemblage

Figure 6.3. The tectonic assemblages and plutonic suites of the Uchi Subprovince, *modified from* the Tectonic Assemblages of Ontario, northern and west-central sheets (*see* Maps 2575 and 2576, map case), and showing principal younging directions and U-Pb age determinations. For locations of U-Pb age determinations, in the Red Lake and Birch–Uchi greenstone belts, *see* the more detailed Figures 6.4 and 6.5; for the Meen–Dempster, Lake St. Joseph and Pickle Lake greenstone belts, *see* Figures 6.6, 6.9 and 6.10. Refer to Figure 6.2 for location of intrusions, greenstone belts and lakes.

submarine plains that characterize, for example, the central part of the Red Lake greenstone belt;

2. chemical sediments deposited at or near the top of dacitic to rhyolitic pyroclastic deposits on the submarine slopes of volcanoes, an example of which is preserved in the northwestern part of the Red Lake greenstone belt;
3. clastic sequences deposited on the flanks of volcanic edifices and derived principally from the erosion of the edifices, as preserved in the northern part of western Lake St. Joseph in the Confederation and St. Joseph assemblages;
4. syntectonic submarine fan and alluvial and/or fluvial deposits, of mixed, volcanic and plutonic provenance, which unconformably overlie the volcanic sequences and could, in some cases, be compared to modern foreland basin sequences or they may be tectonically transported parts of a postulated accretionary prism, which lies to the south of the Uchi Subprovince. The remnants of these basins are preserved in the clastic sedimentary assemblages, such as the Billett assemblage in the Meen–Dempster greenstone belt and other unnamed assemblages along the length of the subprovince, from the Red Lake greenstone belt to the Miminiska–Fort Hope greenstone belt (*see* Figure 6.3).

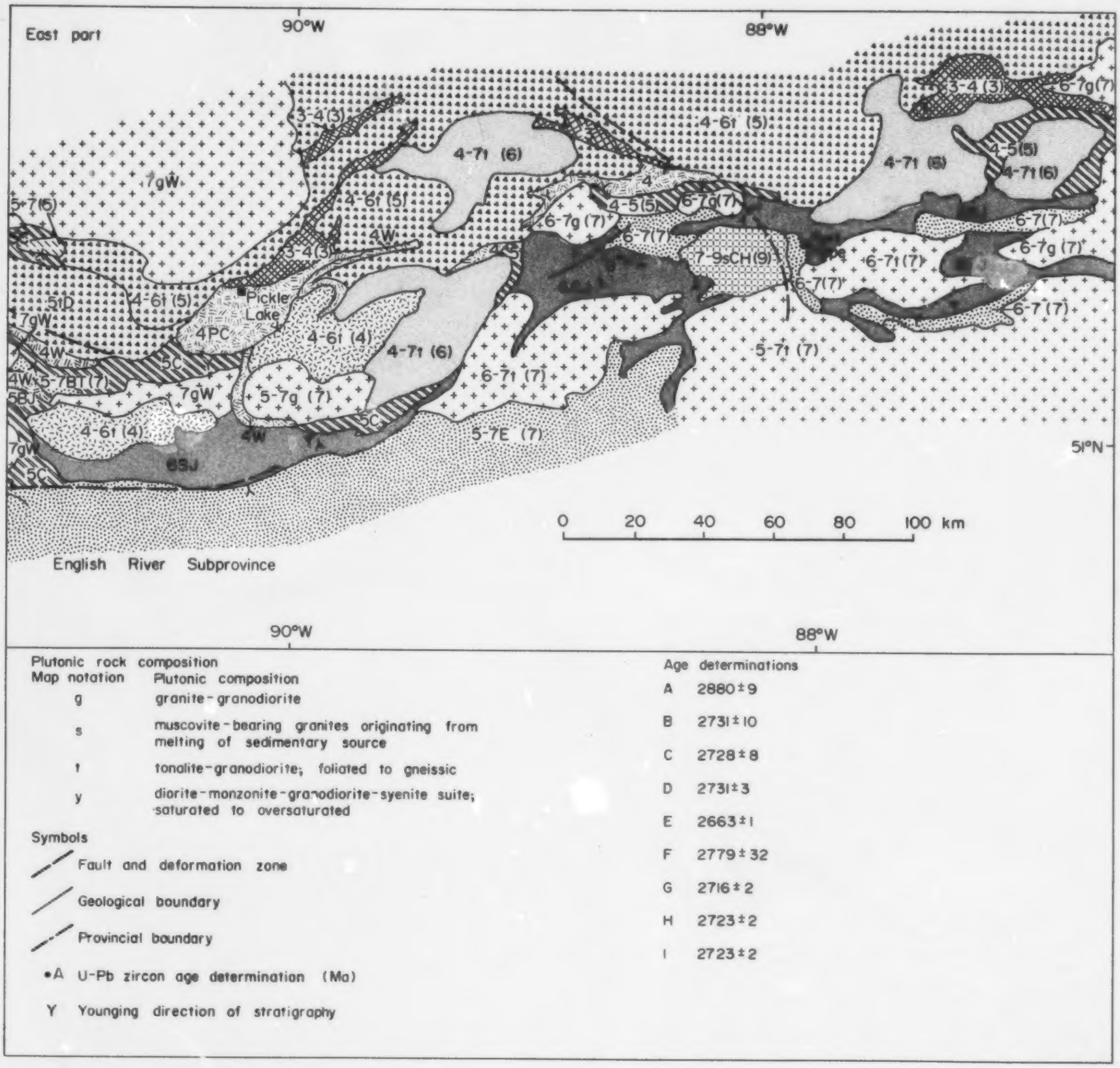

East part
90°W
88°W
3-4(3)
4-6t (5)
4-7t (6)
4-7t (6)
3-4 (3)
6-7g(7)
4-5(5)
4-7t(6)
7gW
5-7(5)
4-6t (5)
4W
3-4(3)
6-7g (7)
4-5(5)
6-7g(7)
6-7(7)
7-9sCH(9)
6-7(7)
6-7t (7)
6-7(7)
6-7g (7)
6-7(7)
5tD
4-6t (5)
Pickle Lake
4PC
4-6t (4)
4-7t (6)
7gW
4W
5C
4W 5-7BT(7)
5BJ
7gW
5-7g (7)
6-7t (7)
5-7t (7)
4-6t (4)
4W
5C
5-7E (7)
7gW
6SJ
5C
51°N
0 20 40 60 80 100 km
English River Subprovince
90°W
88°W
Plutonic rock composition
Map notation Plutonic composition
g granite-granodiorite
s muscovite-bearing granites originating from melting of sedimentary source
t tonalite-granodiorite; foliated to gneissic
y diorite-monzonite-granodiorite-syenite suite; saturated to oversaturated
Symbols
Fault and deformation zone
Geological boundary
Provincial boundary
•A U-Pb zircon age determination (Ma)
Y Younging direction of stratigraphy
Age determinations
A 2880±9
B 2731±10
C 2728±8
D 2731±3
E 2663±1
F 2779±32
G 2716±2
H 2723±2
I 2723±2

Figure 6.3. The tectonic assemblages and plutonic suites of the Uchi Subprovince, *modified from* the Tectonic Assemblages of Ontario, northern and west-central sheets (*see* Maps 2575 and 2576, map case), and showing principal younging directions and U-Pb age determinations. For locations of U-Pb age determinations, in the Red Lake and Birch–Uchi greenstone belts, *see* the more detailed Figures 6.4 and 6.5; for the Meen–Dempster, Lake St. Joseph and Pickle Lake greenstone belts, *see* Figures 6.6, 6.9 and 6.10. Refer to Figure 6.2 for location of intrusions, greenstone belts and lakes.

Summary of Regional Geology

The Uchi Subprovince is a tabular, eastward-trending region of metavolcanic and lesser metasedimentary rocks forming a semicontinuous supracrustal network interweaving around granitoid batholiths and plutons. The long axis of the subprovince trends approximately 080°. In its map pattern, the Uchi Subprovince differs from other volcano-plutonic subprovinces of the Superior Province by its narrower ribbon-like form. Its eastward-trending structural grain differs from the adjacent Sachigo Subprovince, where the structural grain trends southeastward. This contrast becomes less obvious with detailed study. The ribbon-like form of the Uchi Subprovince is accentuated by the presence of the Berens River (plutonic belt) Subprovince and the English River Subprovince, each striking eastward and, respectively, bounding the north and south margins of the Uchi Subprovince (*see* Figure 6.1). The southern boundary with the metasedimentary-plutonic English River Subprovince is the Sydney Lake–Lake St. Joseph Fault (*see* Figure 6.2) and the northern boundary is gradational with the Berens River Subprovince, a domain of plutons and tonalitic gneisses.

The supracrustal rocks of the Uchi Subprovince are informally subdivided into several greenstone belts shown in Figures 6.2, 6.4, 6.5 and 6.6, which include location names to aid in the following descriptions.

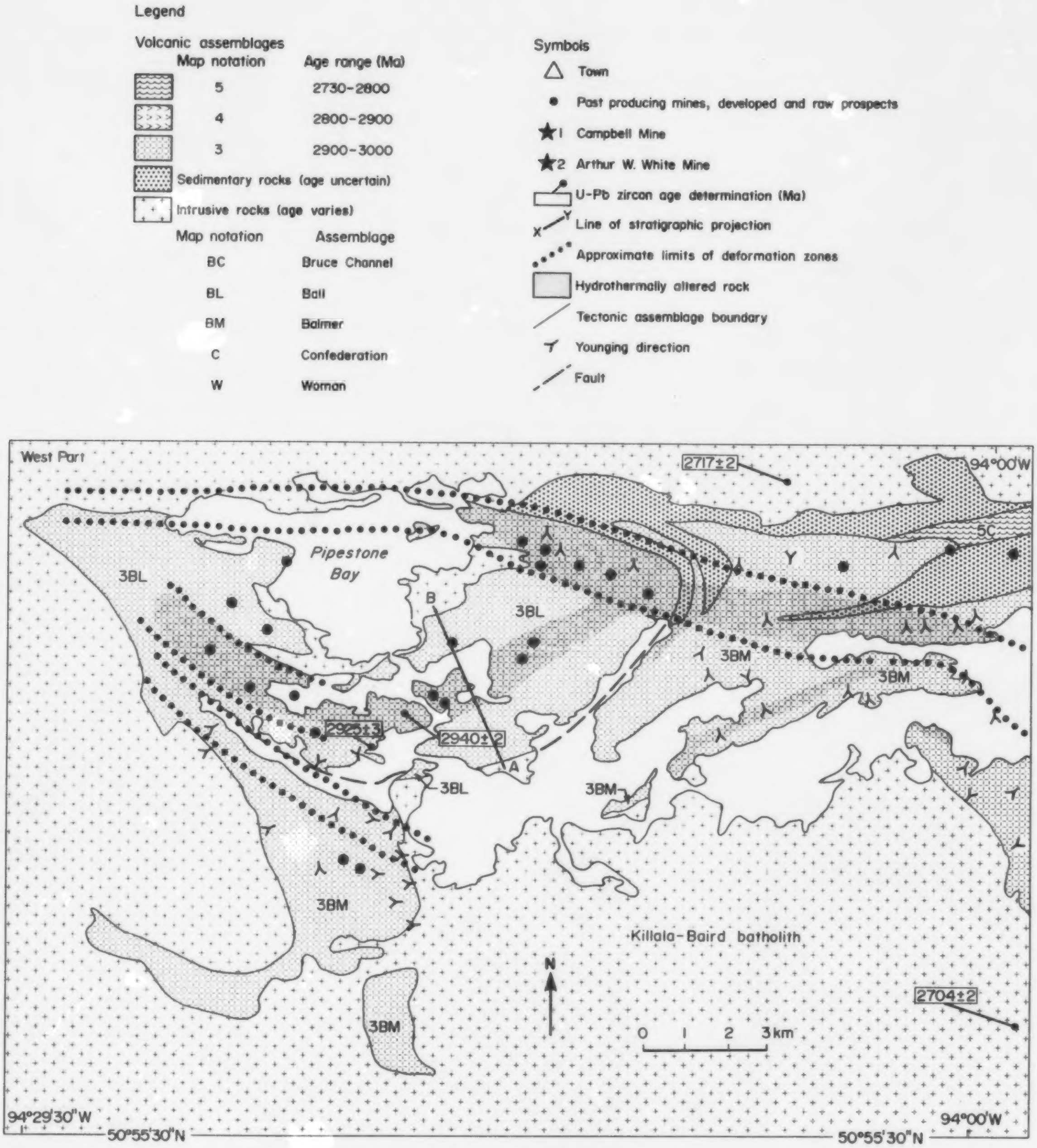

Legend
Volcanic assemblages
Map notation
Age range (Ma)
5
2730-2800
4
2800-2900
3
2900-3000
Sedimentary rocks (age uncertain)
Intrusive rocks (age varies)
Map notation
Assemblage
BC
Bruce Channel
BL
Ball
BM
Balmer
C
Confederation
W
Woman
Symbols
Town
Past producing mines, developed and raw prospects
1 Campbell Mine
2 Arthur W. White Mine
U-Pb zircon age determination (Ma)
Line of stratigraphic projection
X
Y
Approximate limits of deformation zones
Hydrothermally altered rock
Tectonic assemblage boundary
Younging direction
Fault
West Part
2717±2
94°00'W
5C
Y
Pipestone Bay
3BL
B
3BL
3BM
3BM
2925±3
2940±2
A
3BL
3BM
3BM
Killala-Baird batholith
N
0 1 2 3km
2704±2
3BM
94°29'30"W
50°55'30"N
50°55'30"N
94°00'W

Figure 6.4. A map of the tectonic assemblages in the Red Lake greenstone belt, showing critical U-Pb zircon age determinations of volcanic and plutonic rocks, detailed stratigraphic-younging directions, and deformation zones, the limits of which contain a higher than normal abundance of discrete ductile and brittle-ductile faults and shear zones (*modified from* Andrews et al. 1986; Durocher et al. 1991). The boundary between assemblages is typically a fault, which possibly marks the zone along which 1 assemblage was tectonically transported into juxtaposition with the other. Since sedimentary sequences in the northern part of the belt may be associated with an adjacent volcanic assemblage, or allochthonously transported northward from the English River assemblage, they have not been assigned to an assemblage for now. For locations of stratigraphic sections, *see* Figure 6.7.

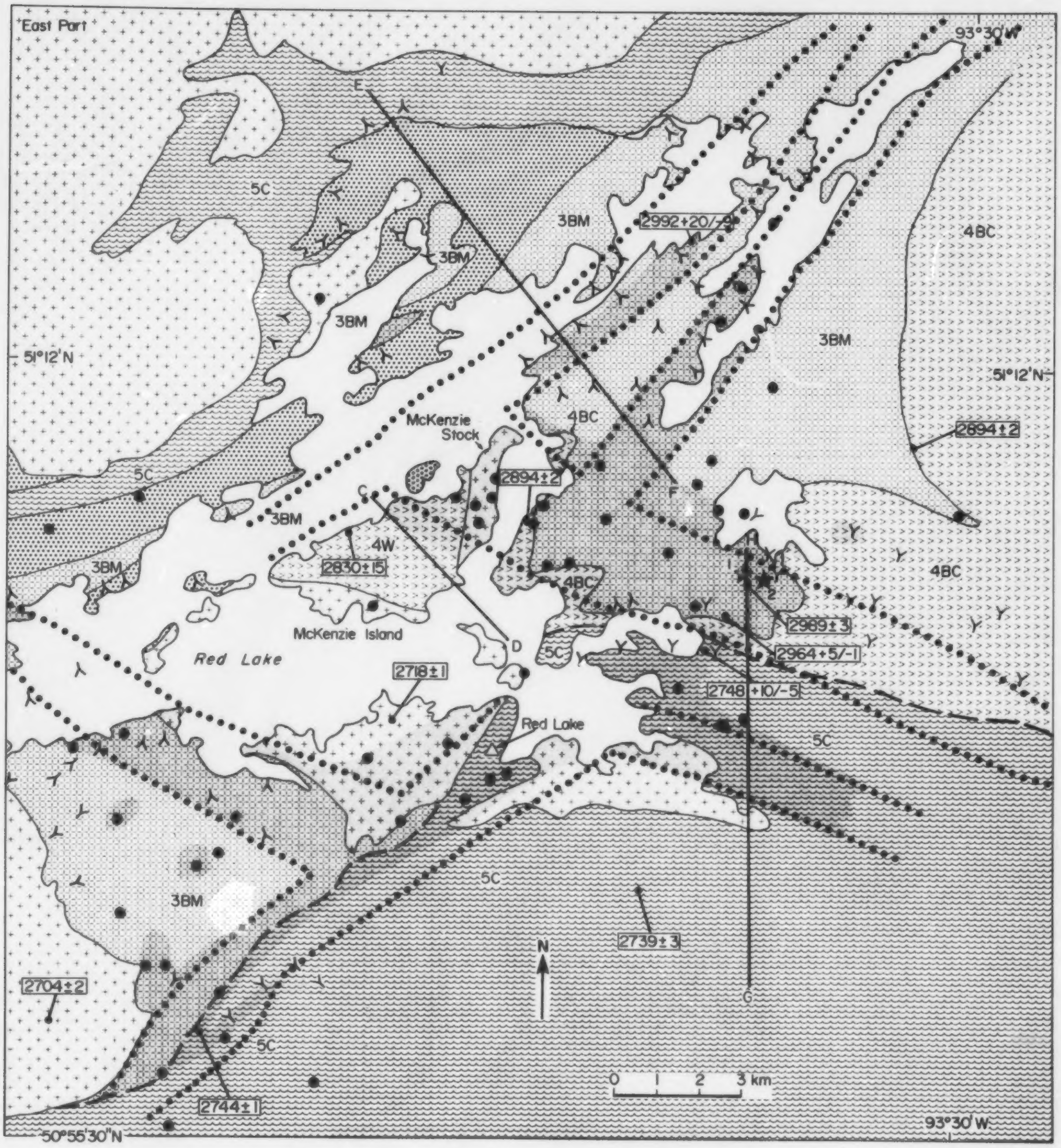

East Part
93°30'W
5C
E
3BM
3BM
3BM
2992+20/-9
4BC
51°12'N
3BM
51°12'N
McKenzie Stock
4BC
2894±2
2894±2
F
5C
C
3BM
4W
2830±15
H
I
2
4BC
4BC
3BM
McKenzie Island
Red Lake
D
2989±3
2964+5/-1
5C
2718±1
2748+10/-5
Red Lake
5C
5C
3BM
2739±3
N
2704±2
G
5C
0 1 2 3 km
2744±1
50°55'30"N
93°30'W

Figure 6.4. A map of the tectonic assemblages in the Red Lake greenstone belt, showing critical U-Pb zircon age determinations of volcanic and plutonic rocks, detailed stratigraphic-younging directions, and deformation zones, the limits of which contain a higher than normal abundance of discrete ductile and brittle-ductile faults and shear zones (*modified from* Andrews et al. 1986; Durocher et al. 1991). The boundary between assemblages is typically a fault, which possibly marks the zone along which 1 assemblage was tectonically transported into juxtaposition with the other. Since sedimentary sequences in the northern part of the belt may be associated with an adjacent volcanic assemblage, or allochthonously transported northward from the English River assemblage, they have not been assigned to an assemblage for now. For locations of stratigraphic sections, *see* Figure 6.7.

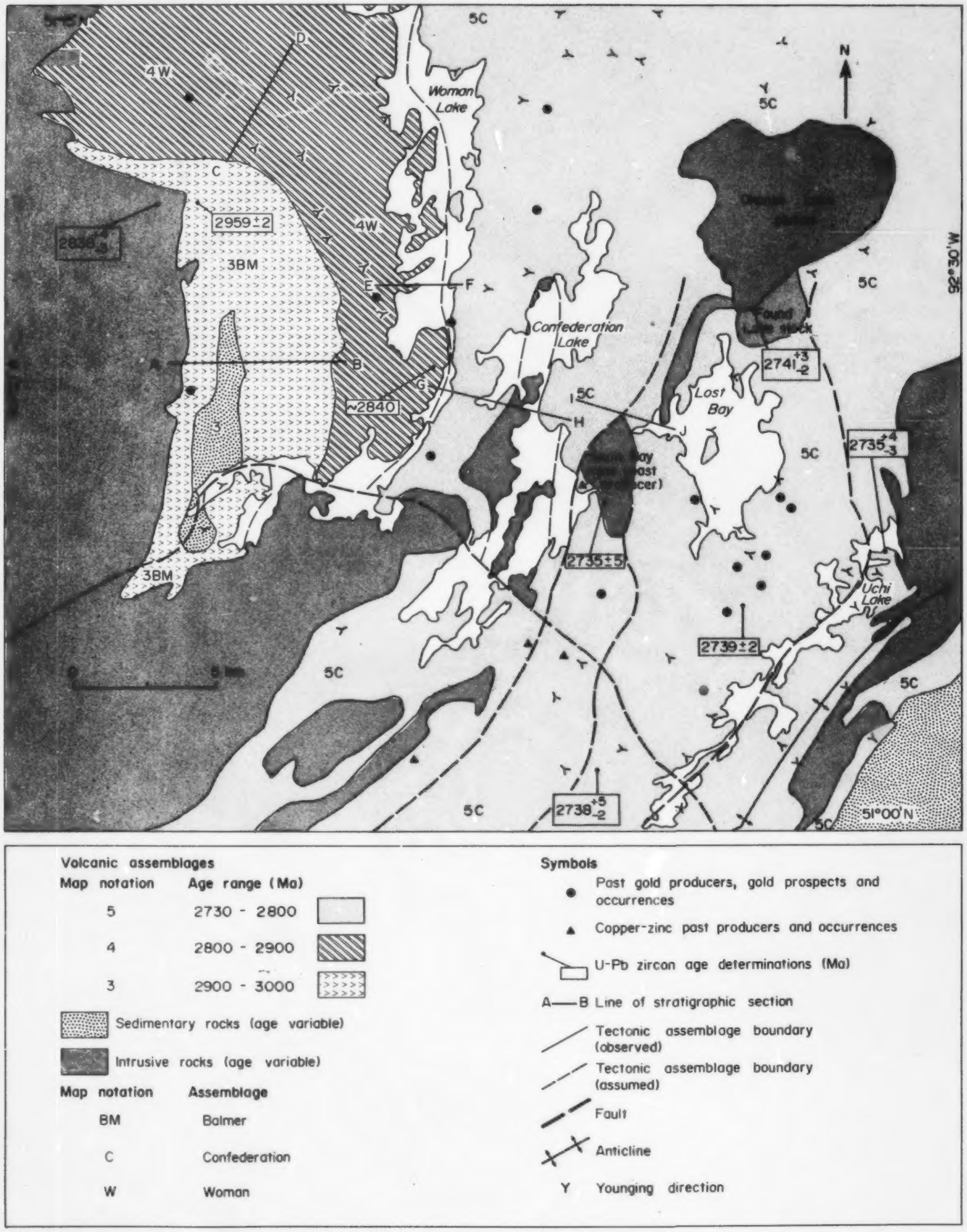

4W
D
C
2959±2
2838
3BM
Woman Lake
5C
N
E
F
A
B
G
~2840
Confederation Lake
5C
I
H
J
Lost Bay
Found Lake stock
2741 +3 -2
2735 +4 -3
2735±5
2739±2
2738 +5 -2
Uchi Lake
3BM
3
0
5 km
51°00'N
92°30'W
Volcanic assemblages
Map notation
Age range (Ma)
5
2730 - 2800
4
2800 - 2900
3
2900 - 3000
Sedimentary rocks (age variable)
Intrusive rocks (age variable)
Map notation
Assemblage
BM
Balmer
C
Confederation
W
Woman
Symbols
Past gold producers, gold prospects and occurrences
Copper-zinc past producers and occurrences
U-Pb zircon age determinations (Ma)
A—B Line of stratigraphic section
Tectonic assemblage boundary (observed)
Tectonic assemblage boundary (assumed)
Fault
Anticline
Y Younging direction

Figure 6.5. A map of tectonic assemblages in the Birch–Uchi greenstone belt, showing critical U-Pb zircon age determinations of volcanic and plutonic rocks, detailed stratigraphic-younging directions and major faults. Boundaries at the top of the Balmer and Woman assemblages may be faulted. For locations of stratigraphic sections, *see* Figure 6.8.

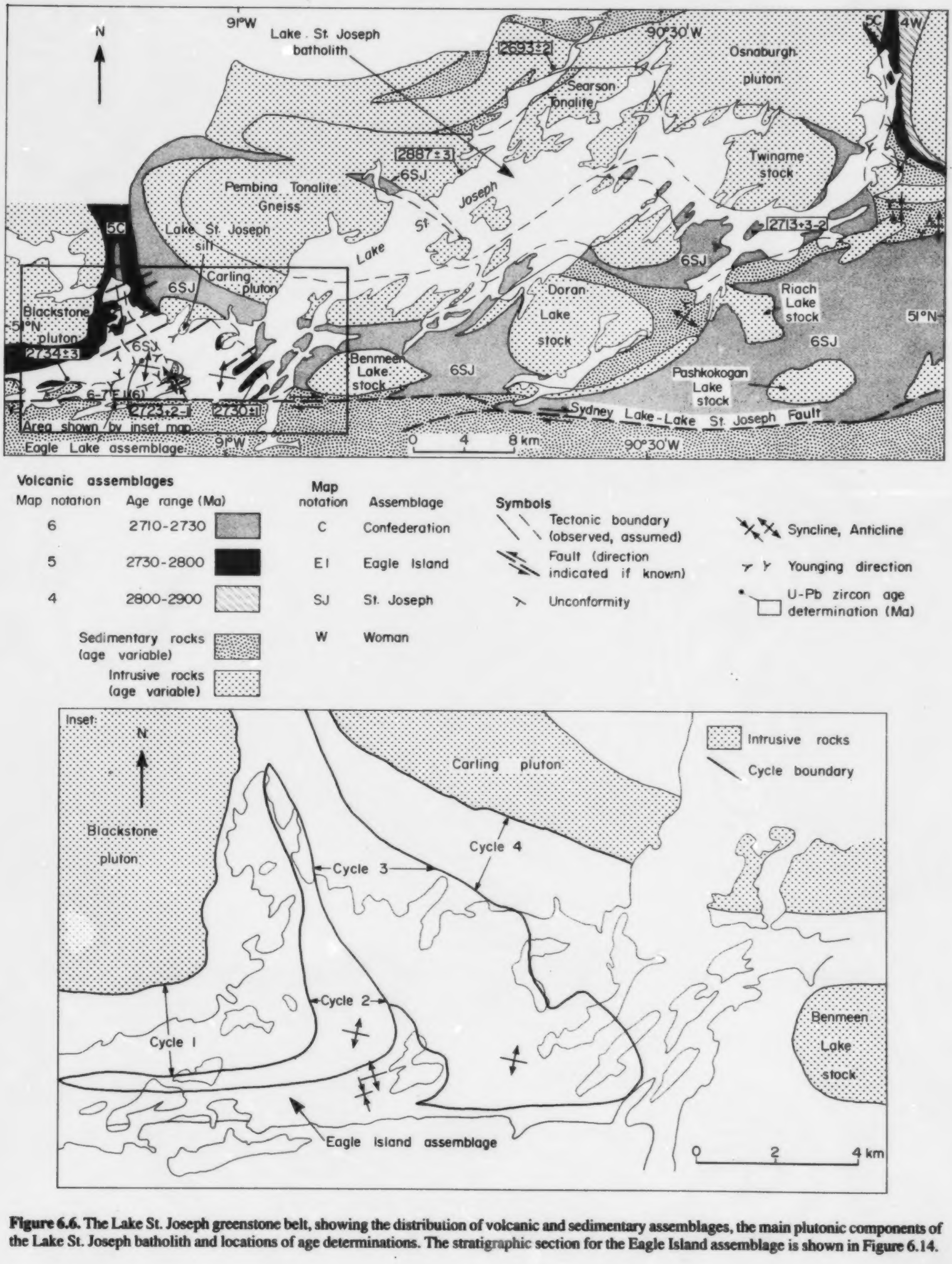

N
91°W
Lake St. Joseph batholith
2693±2
90°30'W
Osnaburgh pluton
5C
4W
Searson Tonalite
2887±3
6SJ
Twiname stock
Pembina Tonalite Gneiss
Lake St. Joseph
5C
Lake St. Joseph sill
Carling pluton
6SJ
2713+3-2
6SJ
Doran Lake stock
Riach Lake stock
51°N
Blackstone pluton
51°N
2734±3
6SJ
6SJ
Benmeen Lake stock
6SJ
6-7EI(6)
2723+2-1
2730±1
Pashkokogan Lake stock
Sydney Lake-Lake St. Joseph Fault
Area shown by inset map
Eagle Lake assemblage
91°W
0 4 8 km
90°30'W
Volcanic assemblages
Map notation
Age range (Ma)
6 2710-2730
5 2730-2800
4 2800-2900
Sedimentary rocks (age variable)
Intrusive rocks (age variable)
Map notation
Assemblage
C Confederation
EI Eagle Island
SJ St. Joseph
W Woman
Symbols
Tectonic boundary (observed, assumed)
Fault (direction indicated if known)
Unconformity
Syncline, Anticline
Younging direction
U-Pb zircon age determination (Ma)
Inset:
N
Blackstone pluton
Carling pluton
Intrusive rocks
Cycle boundary
Cycle 4
Cycle 3
Cycle 2
Cycle 1
Benmeen Lake stock
Eagle Island assemblage
0 2 4 km

Figure 6.6. The Lake St. Joseph greenstone belt, showing the distribution of volcanic and sedimentary assemblages, the main plutonic components of the Lake St. Joseph batholith and locations of age determinations. The stratigraphic section for the Eagle Island assemblage is shown in Figure 6.14.

Exposure of the Uchi Subprovince rocks generally decreases eastward as the blanket of glacial overburden becomes increasingly prominent and the topography flattens. Accordingly, the bedrock exposure in many parts of the Uchi Subprovince is poor and best revealed along lakeshores and rivers. The flat topography of the Uchi Subprovince is interrupted by low hills of glacial moraines, eskers and drumlins elongated to the southwest (*see* Barnett, this volume).

Tectono-Stratigraphic Framework of the Greenstone Belts

MAIN FEATURES

The Uchi Subprovince is characterized by patterns in age distribution, rock types and structure. The most notable patterns as given below are evident from Figure 6.3, Tables 6.1 to 6.7 and OGS Map 2576 (*see* Map 2576, map case) .

Age. The oldest supracrustal rocks are found in the Red Lake greenstone belt, the western flank of the Birch–Uchi greenstone belt and farther west in parts of the Uchi Subprovince in Manitoba (Corfu and Wallace 1986; Corfu and Andrews 1987; Turek et al. 1989; *see* Figure 6.2). The subprovince is volumetrically dominated by volcanic rocks ranging in age from 2730 to 2745 Ma. The youngest volcanic rocks lie south and east of Lake St. Joseph and farther east in the southern Miminiska–Fort Hope greenstone belt. There is an overall sense of younging of volcanic assemblages from north to south across the subprovince, but in detail the pattern is more complex, substantiated by the presence of younger supracrustal rocks on the north side of the Red Lake greenstone belt (*see* Figure 6.3) and north of the Meen–Dempster greenstone belt.

Rock Types. The Uchi Subprovince is characterized by thick sequences of basaltic flows as in the Balmer assemblage in the Red Lake greenstone belt, similar to Phanerozoic oceanic basaltic plains, and mafic to felsic volcanic cycles dominated by bimodal volcanism (basalt to dacite ± rhyolite) as in the Confederation assemblage, present in several greenstone belts similar to Phanerozoic magmatic arcs. An exception is the uppermost volcanic cycle in the central Uchi Subprovince on Lake St. Joseph, which is a thick sequence dominated by andesitic to dacitic pyroclastic flows. Typically, komatiitic flows are rare in much of the Uchi Subprovince, particularly outside of the older volcanic sequences (greater than 2900 Ma) known especially in the western part of the subprovince.

Structure. The Uchi Subprovince forms the eastward-trending southern flank of a tectonic block that includes the Berens River and Sachigo subprovinces as defined by Card and Ciesielski (1986). The southern boundary is sharply defined by the east-trending Sydney Lake–Lake St. Joseph Fault. Supracrustal belts and the structural grain of intrusions generally trend to the east in the southern part of the Uchi Subprovince. Northern parts of the Uchi Subprovince feature northwest- and northeast-trending belts that, from aeromagnetic and field relationships, appear

Table 6.1. Assemblage of the Bee Lake Belt.

Name	Rock Types, Primary Structures (described from base to top where known)	Contacts	Interpreted Environment	Age (Ma)	References
Confederation	–massive and minor pillowed basalt flows –dacite, tuff breccia and pyroclastic breccia grading to tuff farther east –arkose and arkose pebble conglomerate with graded bedding, scour structures; conglomerate is heterolithic with volcanic, sedimentary, granitic clasts; arkose of quartz and plagioclase; banded, magnetite iron formation and chloritic phyllite units	–part of Gem Lake Subgroup of Rice Lake greenstone belt in Manitoba –Gem Lake Subgroup thrust fault contact with and overlying anticlinal, Bidou Lake Subgroup; stratigraphic top directions of each subgroup face toward the other subgroup –transected and folded at Sydney Lake–Lake St. Joseph Fault	arc volcanism	<2730? comparable to or younger than units of Rice Lake greenstone belt in Manitoba	Shklanka (1967) Weber (1971a) McRitchie and Weber (1971b) Turek et al. (1989)

Table 6.2. Assemblages of the Red Lake greenstone belt.

Name	Rock Types, Primary Structures (described from base to top where known)	Contacts	Interpreted Environment	Age (Ma)	References
Balmer	– massive to pillowed basaltic tholeiite and komatiite flows with interbeds of banded magnetite iron formation – minor tholeiitic rhyolite, dominated by pyroclastic breccia to fine-grained tuff	– dominantly northward-facing assemblage – in fault contact with southern Confederation assemblage and Ball assemblage – uncertain relation to unnamed sedimentary assemblage	oceanic volcanism, local shallow submarine	2992^{+20}_{-9} rhyolite ash flow breccia 2989 ±3 rhyolite flow >2982 (min) rhyolite flow 2964^{+5}_{-1} rhyolite flow	Pirie (1981, 1982) Wallace et al. (1986) Corfu and Wallace (1986) Corfu and Andrews (1987) Durocher et al. (1991)
Ball	– dominantly calc-alkalic sequence of interlayered massive to pillowed mafic volcanic flows and more prominent intermediate to felsic flows and pyroclastic lapilli tuff and tuff – dolomitic marble-chert beds, near stratigraphic top of assemblage, contains stromatolites	– fault contact in lake with Balmer assemblage; bedding of Ball and Balmer assemblages face toward each other	subaqueous and subaerial arc volcanism; local shallow submarine	2940±2 rhyolite lapilli tuff 2925±3 rhyolite flow	Riley (1972, 1975) Corfu and Wallace (1986) Hofmann et al. (1985)
Bruce Channel	– dominantly massive and pillowed basalt, poorly exposed with cap of minor felsic, tuffaceous rocks, wacke-sandstone and magnetite iron formation	– tectonically underlies the older Balmer assemblage and contact is folded near Balmer Lake – in fault contact with Confederation assemblage – undetermined contact with Woman assemblage—possible unconformity	arc volcanism	2894 ±2 rhyolite tuff	Corfu and Wallace (1986)
Woman	– dominantly felsic to intermediate pyroclastic rocks, typically heterolithic clasts in lapillistone and lapilli tuff	– assemblage is a small window or sliver surrounded by Red Lake and bordered by faults; intruded by 2718 million-year-old Dome stock – relation to other assemblage not known	arc volcanism	2830± 15 heterolithic lapilli tuff	Corfu and Wallace (1986) Wallace et al. (1986)
Confederation	– dominantly calc-alkalic units in southern part: complexly interbedded and interdigitated pillowed, variolitic basalt and andesite flows with plagioclase or amphibole phenocrysts – locally thick deposits of andesitic breccia and lapillistone – dacitic to andesitic flows and pyroclastic lapillistone and breccia – interbedded rhyolitic flows, beds of tuff, lapillistone and pyroclastic breccia – northern part: dominantly dacitic and rhyolitic pyroclastic deposits	– Flat Lake–Howey Bay deformation zone marks boundary of southern Confederation assemblage with Balmer assemblage; part of boundary intruded and stitched by 2718 million-year-old Dome Stock – nature of contact between northern Confederation assemblage and both Balmer and unnamed sedimentary assemblages is undetermined; northern assemblage interpreted as allochthonous, transported northward and separated from main southern part; 2717 million-year-old external granodiorite pluton intrudes north	arc volcanism	Southern: 2748^{+10}_{-5} heterolithic lapilli tuff 2744 ±1 lapilli crystal tuff 2739 ±3 rhyolite crystal tuff Northern: 2733 ± 2 rhyodacite tuff	Pirie (1982) Corfu and Wallace (1986)

Table 6.3. Assemblages of the Birch–Uchi greenstone belt.

Name	Rock Types, Primary Structures (described from base to top where known)	Contacts	Interpreted Environment	Age (Ma)	References
Balmer	Cycle I of Thurston (1985a) up to 4800 m thick of formations: Fm. A: subaqueous mafic flows, massive and pillowed and amphibole-phyric, with minor intermediate tuff and lapilli tuff; up to 2500 m thick Fm. B: wedge of mixed wacke, argillite and intermediate lapilli tuff, and minor felsic tuff; up to 1500 m thick Fm. C: small unit of felsic lapilli tuff and minor crystal tuff, interfingering at the top of Fm. A; up to 320 m thick Fm. D: shallow water marble up to 100 m thick and local chert and sulphide facies iron formation 60 to 70 m thick	– intruded by Narrow Lake tonalite (2838 Ma) of Trout Lake batholith; undetermined disconformable contact with Woman assemblage, but tonalite is synchronous with Woman assemblage volcanism, and its intrusion into Balmer assemblage may indicate original extrusion of Woman volcanic rocks upon Balmer assemblage; subsequent 2805 million-year-old tonalite of Trout Lake batholith crosscuts Balmer–Woman assemblages boundary	subaqueous oceanic volcanism with local shallow subaqueous volcanism and shallow water chemical sedimentation	2959 ±2 rhyolite crystal lithic tuff (Fm. C)	Cycle I of Thurston (1985a) Nunes and Thurston (1980)
Woman	Cycle II of Thuston (1985a): Fm. E: mafic and intermediate subaqueous flows with minor intermediate and felsic tuff, cherty interflow units, and oxide facies iron formation; up to 3600 m thick Fm. G: proximal to distal facies of thickly bedded, massive flows comprising intermediate lapilli tuff and tuff; up to 2800 m thick Fm. H: subaerial ignimbrite sheets and subaqueous ash flows of felsic lapilli tuff, and lesser tuff, tuff breccia and porphyritic flows; up to 720 m thick Fm. J: shallow submarine stromatolite-bearing marble plus chert and oxide facies iron formation; up to 150 m thick		subaqueous shield volcanism with subaerial volcanism and shallow water sedimentation in arc setting	2840 (est.) amygdular rhyolite flow (Fm. 8)	Cycle II of Thurston (1985a) Corfu (unpublished)
Confederation	Cycle III of Thurston (1985a): Fm. K: mafic to intermediate pillowed flows with hyaloclastite debris, massive variolitic flows and interflow sediment; up to 1670 m thick Fm. L: proximal, shallow water, intermediate tuff and lapilli tuff with minor intermediate flows and tuff breccia; up to 2570 m thick Fm. M: mainly within a fault-bounded block, unit (up to 200 m thick) comprising felsic flows, tuff and intermediate flows of proximal to vent facies breccias from phreatic explosive activity; endogenous spherulitic partly welded tuffs, lapilli tuffs, rhyolite tuff; – quartz-feldspar porphyry dome – fault-bounded graben of Caldera sequence of rhyolite tuff, flows	– central axis of assemblage marked by fault toward which stratigraphic units face; axis is probably not a syncline – undetermined nature of east contact with underlying, unnamed sedimentary assemblage – contact on west with Woman assemblage appears conformable, but may be a fault and is marked by local shear zones cutting obliquely across the boundary	arc volcanism	2739 ±2 rhyolitic vitric tuff 2738^{+5}_{-2} rhyolite flow 2735^{+4}_{-3} rhyolite tuff	Thurston et al. (1978) Cycle III of Thurston (1985a) Nunes and Thurston (1980) Noble (1989) Noble et al. (1989)

Table 6.4. Meen–Dempster and Pickle Lake greenstone belts.

Name	Rock Types, Primary Structures (described from base to top where known)	Contacts	Interpreted Environment	Age (Ma)	References
Unnamed assemblage (North Pickle Lake greenstone belt)	– massive and locally pillowed basalt flows with banded, magnetite iron formation – minor dacite tuff interbeds	– inferred fault contact with Pickle Crow assemblage from angular discordance of geophysical conductors in each assemblage – (?) assemblage may correlate with McGruer assemblage of North Caribou greenstone belt	oceanic volcanism	2.9 Ga (est.)	Stott et al. (1989a, 1989b)
Pickle Crow	– pillowed tholeiitic basalt flows, intruded by quartz-porphyry sills and overlain by dacitic lapillistone, lapilli tuff, tuff and thick beds of fine-grained tuff, possibly of ash flows – pillowed to massive basalt flows, with banded, magnetite iron formation interbeds	– undetermined nature of contact with Woman assemblage – (?) possible correlation with Bruce Channel assemblage of Red Lake greenstone belt and with unnamed assemblage in north Miminiska–Fort Hope greenstone belt	arc volcanism	2892^{+10}_{-2} quartz porphyry	Stott et al. (1986) Corfu (unpublished data) Corfu and Stott (1989)
Woman	– massive and minor pillowed tholeiitic basalt flows with banded magnetite iron formation interbeds; unit at least 3 km wide interlayered with units of dacitic, quartz-phyric and fine-grained tuff, probably distal ash flows; interbedded with chloritic mudstone and chert – local proximal ash flows overlain by ferruginous chert beds, and local thin quartz-marble unit	–complex and undetermined bedding-parallel juxtaposition with Confederation assemblage; inferred fault contact – intrusive contact with 2713 million-year-old Second Loon pluton and 2740 million-year-old Ochig Lake pluton	oceanic volcanism with local subaqueous to subaerial arc sequence rhyodacitic	2841^{+5}_{-2} dacitic quartz-phyric tuff 2836 (est.) dacitic quartz-phyric tuff 2825 ±2 dacitic lapilli tuff 2816 ±1 dacitic tuff 2805^{+10}_{-5} tuff	Corfu (unpublished data)
Confederation	– vesicular, pillowed and massive tholeiitic basalt flows with flow top breccia, 2000 m thick – overlain by distal fine-grained dacitic ash flows and resedimented tuff, and capped by volcanogenic sedimentary rocks as a typically 1000 m thick unit	– probable fault contact; assemblage base locally oblique to underlying Woman assemblage – unconformably overlain after internal tectonic stacking, by sedimentary basin of Billett assemblage	oceanic volcanism	2740 ±1 2744 ±2	Corfu (unpublished data) Corfu and Stott (1989)

Table 6.5. Assemblages of the Lake St. Joseph greenstone belt.

Name	Rock Types, Primary Structures (described from base to top where known)	Contacts	Interpreted Environment	Age (Ma)	References
Confederation	– tholeiitic basalt flows with massive to pillowed facies progression in each flow unit; iron carbonate amygdules increase in abundance upward in stratigraphic section – rhyolite-rhyodacite pyroclastic units of clast-supported flow breccias, pyroclastic flows with coarse lithic clasts and scarce rounded, massive sulphide clasts; abundant basalt dikes; pervasive iron carbonatization	– base is intruded by 2701 Ma Blackstone pluton – Confederation assemblage disconformably overlain by St. Joseph assemblage, apparently sharing same volcanic vent system with St. Joseph assemblage – sequence of Confederation and overlying St. Joseph assemblages is repeated by unobserved thrust fault	arc volcanism, proximal and vent facies	2734 ± 3 2730 ± 1	Clifford (1969) Berger (1981) Stott et al. (1987a, 1987b) Corfu and Stott (1989) Corfu (unpublished data)
St. Joseph	– hydrothermally altered, bleached basalt flows overlain by dominant vent facies sequence of massive rhyolite quartz-feldspar phyric flow, phreatic breccia, overlain by a suite of pyroclastic flow deposits; clast sizes diminish on flanks of volcano away from vent area to distal dacitic to andesitic tuff and volcaniclastic sedimentary rock – uppermost pyroclastic flow deposits 1000 m thick and traceable in a unit over 60 km long, of dacitic, lithic tuff breccia, vesicular tuff breccia, and more voluminous dacitic to andesitic lapilli tuff, increasingly more mafic up-section	– hiatus in volcanic activity between Confederation and St. Joseph assemblages – unconformably overlain by Eagle Island assemblage	arc volcanism, proximal and vent facies	273^{+2}_{-1} 2713^{+3}_{-2} andesite tuff	Clifford (1969) Berger (1981) Stott et al. (1987a, 1987b) Corfu and Stott (1989) Corfu (unpublished data)
Billett	– turbiditic AC wacke and banded magnetite iron formation interbeds – local interbeds of graded sandstone and conglomerate on north margin; trondhjemite cobble and pebble clasts derived from exhumed Dobie Lake pluton	– angular unconformity with and overlying tectonically juxtaposed Woman and Confederation assemblages – cut by 2693 Ma Osnaburgh pluton – assemblage may be correlative with unnamed sedimentary assemblages in north Birch–Uchi greenstone belt and central Miminiska–Fort Hope greenstone belt	late orogenic basin	>2699	Stott and Wilson (1986b) <2740 (2710 est.)
Bamaji	>700 m thick assemblage of: – basal conglomerate overlying North Bamaji pluton, the source of cobble clasts – thin dacite to latitic pyroclastic unit of crystal lithic tuff and lapilli tuff beds – interlayered with wacke units	– unconformably overlies 2805 Ma Woman assemblage – erosive unconformity with conglomerate overlying exhumed 2806 Ma North Bamaji pluton	late orogenic wrench fault-related or pull-apart basin	2781^{+4}_{-3} felsic tuff	Wallace (1985) Scharer (1989)

Table 6.6. Assemblage of the Lang Lake belt.

Name	Rock Types, Primary Structures (described from base to top where known)	Contacts	Interpreted Environment	Age (Ma)	References
Confederation	– massive to pillowed basalt flows with interflow, banded, magnetite iron formation units that continue into wacke sedimentary sequence at east end of belt – fine-grained tuff to pyroclastic breccia and quartz porphyry dikes – apparent syncline interpreted along central east-west axis of belt	– isolated belt; apparently coeval with Dobie Lake pluton	arc volcanism	2749^{+8}_{-3}	Fenwick (1970, 1971) Fenwick and Srivastava (1972)

Table 6.7. Assemblages of the Miminiska–Fort Hope greenstone belt.

Name	Rock Types, Primary Structures (described from base to top where known)	Contacts	Interpreted Environment	Age (Ma)	References
St. Joseph	– tholeiitic and calc-alkalic basaltic flows, pillowed and massive with low vesicularity, 7000 to 10 000 m wide – overlain by calc-alkalic andesite, dacite and rhyolite pyroclastic deposits of dominantly tuff and lapilli tuff	– probable fault contact with adjacent older unnamed assemblages to the north in Keezhik Lake area	arc volcanism	2716 ±2 rhyolite tuff 2723 ±2 rhyolite tuff	Wallace (1981a) Corfu (unpublished data) Corfu and Stott (1989)
Unnamed sedimentary assemblage (Miminiska Lake)	– sedimentary beds of medial to distal turbidite wacke with interbeds of banded, magnetite iron formation – thin (600 m wide) conglomerate unit at top of sequence, containing variety of volcanic, iron formation and granitoid clasts	– unconformably overlies St. Joseph assemblage and older unnamed volcanic assemblage in the north half of the belt	late orogenic basin	2710–2700 (est.)	Wallace (1981a)
Older unnamed assemblages	– possibly 2 or more assemblages in Keezhik Lake area, as continuations of strata from Pickle Lake greenstone belt – oldest in north, dominated by massive to pillowed basalt with banded, magnetite iron formation interflow units – younger unnamed assemblage on Keezhik Lake of basalt flows and overlying dacitic lapilli tuff with associated quartz porphyry stock	– stratigraphic magnetic conductors in older and younger assemblage converge at low angle across apparent fault – older assemblage may correlate with McGruer assemblage of North Caribou greenstone belt – younger assemblage may correlate with Pickle Crow assemblage of Pickle Lake greenstone belt	oceanic volcanism in inferred oldest assemblage	2.8–3.0 Ga (est.)	Thurston and Carter (1970)

to discontinuously merge into greenstone belts of the southern Sachigo Subprovince (Stone 1990). For example, a chain of supracrustal slivers and the structural grain of tona-litic gneiss all trend northward from the Red Lake greenstone belt, across the gradational Berens River–Uchi subprovince boundary. Also, elongate aeromagnetic anomalies northeast of Pickle Lake and northwest of Keezhik Lake extend across the Berens River plutonic belt and may include finger-like supracrustal remnants (*see* Map 2542, map case).

GREENSTONE BELTS

The anastomosing array of greenstone belts is a semicontinuous supracrustal network. Mapping at various scales in the past 20 years has subdivided major parts of the belts into volcanic cycles that can be traced for 10s of kilometres along strike. More recently, geochronological investigations by several workers at the Ontario Geological Survey and the Royal Ontario Museum (e.g., Nunes and Thurston 1980; Corfu and Wallace 1986; Corfu and Stott 1989) have, in conjunction with aeromagnetic interpretations and stratigraphic mapping, permitted long-distance correlations of volcanic strata along the length of the subprovince. As shown in the map of Tectonic Assemblages of Ontario (*see* Map 2576, map case) and in Figure 6.3, these correlations in turn permitted us to subdivide the volcanic and sedimentary strata into tectonic assemblages. The principal characteristics of these assemblages for each of the major greenstone belts are summarized in Tables 6.1 to 6.7.

In the following descriptions of assemblages, reference will be made to individual volcanic belts, labelled on Figure 6.2 and used as geographic references, although the belts in most cases are part of a continuum of supracrustal rocks extending the length of the Uchi Subprovince.

BEE LAKE BELT

The Bee Lake belt as described by Shklanka (1967), is the smaller Ontario part of the better known Rice Lake greenstone belt (*see* Figure 6.2) in Manitoba (McRitchie and Weber 1971a). The Rice Lake greenstone belt is composed of the Rice Lake Group of volcanic and sedimentary rocks, subdivided (Weber 1971a) mainly into the Bidou Lake and younger Gem Lake subgroups. These rocks are treated as part of the Confederation assemblage, but a separate northern part of the Rice Lake greenstone belt, near Wallace Lake (*see* Figure 6.2), contains rock types similar to the older 2.9 to 3.0 billion-year-old assemblages in the Red Lake greenstone belt (W. Weber, Manitoba Geological Services Branch, personal communication, 1989) and is included as an unnamed unit with these older rocks (*see* Figure 6.3).

Confederation Assemblage

The Confederation assemblage dominates the southern half of the Uchi Subprovince from Rice Lake in Manitoba to Lake St. Joseph in Ontario (*see* Figure 6.3). It includes rocks ranging in age from 2750 to 2730 Ma, with the younger units of this assemblage typically lying south of the older units.

There is comprehensive documentation of the well-exposed main mass of the belt in Manitoba (Stockwell and Lord 1939; McRitchie and Weber 1971a; Brommecker et al. 1989). Rocks of the Bee Lake belt (*see* Figure 6.2) are tentatively assigned to the Confederation assemblage on the basis of age, and treated as part of the Rice Lake Group. However, in the Rice Lake greenstone belt, rocks of the Gem Lake Subgroup structurally overlie the 2730 million-year-old, anticlinal Bidou Lake Subgroup along a folded boundary, as mapped by Weber (1971b) near the Manitoba–Ontario border, and may be younger or older. Drawing from these structural-stratigraphic studies in the Rice Lake greenstone belt, one can infer from the map pattern that the southeastern extension of this belt into Ontario is a continuation of the Gem Lake Subgroup of the Rice Lake Group described by Weber (1971a, 1971b). Only felsic to intermediate units of the Bidou Lake Subgroup, lying to the west of the Gem Lake Subgroup, have been dated, with consistent magmatic ages of about 2730 Ma (Turek et al. 1989).

The Bee Lake belt is composed of a folded sequence of basaltic flows, intermediate to felsic pyroclastic deposits and an associated sedimentary apron (Shklanka 1967). A summary description of the rocks in the Bee Lake belt is given in Table 6.1. Basaltic flows are generally massive with minor pillowed parts; the rocks are dominantly amphibolitic with greenschist facies equivalents restricted to the centre of the belt. The central part of the belt in Ontario is dominated by proximal, dacitic tuff breccia and pyroclastic breccia; the eastern extremities and the northern and southern flanks of this unit comprise more distal tuff deposits along with arkosic metasedimentary rocks of the Eden Lake area. The central syncline exposes an oblique cross section from distal tuff deposits and clastic, felsic volcanic-derived metasedimentary rocks in the east to more proximal dacite deposits farther west along the belt axis leading to rhyolitic vent deposits on Gem Lake in Manitoba (Weber 1971a). The Eden Lake sedimentary rocks are probably derived from the erosion of this volcanic edifice.

Arkose and arkosic pebble conglomerate dominate the Eden Lake clastic sedimentary rocks at the east end of the belt. Conglomerate clasts are rounded to subrounded, up to cobble and boulder size, and include fragments of intermediate to felsic volcanic, porphyry, iron formation, chert and granitic compositions. Within the Eden Lake sequence is a unit of granitic boulder breccia, 5 km long and up to 450 m thick; the matrix is arkosic and the angular, granitic clasts, up to 60 cm long, reflect a setting rather close to an exhumed plutonic source.

The Bee Lake belt possesses a few primary structures (basaltic pillow shapes and graded bedding in tuff and sedimentary rocks) that Shklanka (1967) used to determine top directions; these, plus iron formation units in the arkosic sedimentary rocks on Eden Lake, and the regional configuration of lithologic units support the present conclusion that the stratigraphy is upside down and therefore, the main part of the belt forms an antiformal syncline that plunges southeastward.

RED LAKE GREENSTONE BELT

The earliest comprehensive description of rock units and mineral deposits and the first analysis of the structural geology of mineral deposits in the Red Lake greenstone belt (*see* Figure 6.2) were reported by Horwood (1945). More recent work encompassed detailed mapping of township areas, as summarized by Wallace et al. (1986), and studies of major shear zones and their relationship to gold deposits by Andrews et al. (1986) and Wilson et al. (1984).

Subdivisions of the Red Lake greenstone belt by Pirie (1981) preceded the later regional stratigraphic correlations made by Wallace et al. (1986) and Corfu and Wallace (1986). The authors of this chapter extended this correlation program to encompass the rest of the Uchi Subprovince.

Balmer Assemblage (Volcanic)

The Balmer assemblage (*see* Figure 6.3) is named for the northern half of Balmer Township, which contains major gold deposits in the Red Lake greenstone belt. The Balmer assemblage incorporates the Lower Mafic Sequence of Pirie (1981, 1982) and Cycle I of Wallace et al. (1986), and forms about 50% of the Red Lake greenstone belt. This assemblage forms the central core of the Red Lake greenstone belt as shown in Figures 6.3 and 6.4. The westernmost part of the Birch–Uchi greenstone belt (*see* Figure 6.2), called Cycle 1 by Pryslak (1971a, 1971b) and Thurston (1985a, 1985b), is interpreted here to form the easternmost extent of the Balmer assemblage, although physically separate from the main mass of this assemblage in the Red Lake greenstone belt. Slivers of amphibolite and clastic sediment form a discontinuous supracrustal ribbon extending northward from the Red Lake greenstone belt into the Berens River Subprovince and along strike with the McInnes Lake greenstone belt (*see* Thurston, Osmani et al., this volume). These supracrustal slivers may be part of a continuation of the Balmer and Ball assemblages beyond the main mass of preserved supracrustal rock forming the Red Lake greenstone belt. Similarly, the northernmost part of the Rice Lake greenstone belt in Manitoba, which was named the Wallace Lake Subgroup of the Rice Lake Group (McRitchie 1971), is regionally on strike with and may correlate with the Balmer or Ball assemblages; if such correlation can be demonstrated, then the Wallace Lake Subgroup would be older than the 2730 million-year-old Rice Lake Group. The Wallace Lake Subgroup rocks (*see* Figure 6.2) compare closely with units of both the Balmer and Ball assemblages, by the predominance of basaltic tholeiite and komatiite, and the presence of quartz arenites and carbonate beds (W. Weber, Manitoba Geological Services Branch, personal communication, 1989).

The Balmer assemblage is composed of the oldest volcanic rocks in the Uchi Subprovince. The ages of volcanism in this assemblage range from about 2958 to 2992 Ma (Corfu and Wallace 1986; Corfu and Andrews 1987). The oldest rhyolitic flows and ash-flow breccia occur in the eastern part of the belt near the Campbell and Arthur W. White mines (*see* Figure 6.2). These felsic volcanic units are comparatively thinner than younger (ca. 2925 to 2940 Ma) Ball assemblage units in the western part of the Red Lake greenstone belt, in the vicinity of Pipestone Bay (*see* Figure 6.4). There are significant age differences among volcanic units in the Balmer assemblage of the eastern Red Lake greenstone belt, probably reflecting a tectonic juxtaposition of strata, and thereby making stratigraphic correlations a challenge for future detailed mapping.

The Balmer assemblage principally features basaltic tholeiite and komatiite lava flows with interbeds of magnetite-quartz iron formation. Rhyolitic pyroclastic breccia to fine-grained tuff occur as comparatively thin units; in places, units possess flattened pumice and welded textures typical of ash-flow deposits. The chemical sediments are lithostratigraphic markers that can be traced in some cases for several 10s of kilometres (Wallace et al. 1986).

The Balmer assemblage separates the Confederation assemblage, which forms the north and south flanks of this belt. This geometric relationship is based on stratigraphic top determinations (*see* Figure 6.4) and differences in the stratigraphic character of the 2 assemblages as illustrated in Figure 6.7. It is also reflected by contrasts in lithogeochemistry and aeromagnetic patterns first recognized by Pirie (1981). The boundary of the Balmer assemblage with the Confederation assemblage along the south flank of the belt is faulted. The Confederation assemblage along the north flank of the belt has probably been tectonically transported northward from the main mass of the Confederation assemblage, but no clear fault boundary along the north flank with the Balmer assemblage has been identified. The boundary relationship between the Balmer assemblage and an unnamed sedimentary assemblage in the northern part of the belt is undetermined.

The boundary between the Balmer and Bruce Channel assemblages is an inferred fault since the younger Bruce Channel assemblage underlies the Balmer assemblage and is therefore stratigraphically out of place. A fault between the Balmer and Ball assemblages is also inferred from oppositely facing stratigraphic younging directions across the boundary.

Ball Assemblage (Volcanic)

Occupying the northwesternmost corner of the Red Lake greenstone belt is the Ball assemblage (*see* Figure 6.3). It is elliptical in plan (up to 16 km long and 7 km wide) and composed of mafic volcanic flows interlayered with intermediate to felsic pyroclastic units. Close to the stratigraphic top are dolomitic marble-chert beds, which partly encircle this assemblage and lie on strike with chert-magnetite iron formation beds (Riley 1972, 1975). The carbonate beds contain stromatolites (Hofmann et al. 1985), one of the few documented localities for stromatolites in the Superior Province.

This assemblage formed the main part of Pirie's (1981) "Ball calc-alkalic sequence". U-Pb zircon ages have been determined (Corfu and Wallace 1986) for rhyolitic lapilli tuff and a rhyolitic flow immediately below and above stromatolitic marble beds in the upper part of the assemblage. These respective ages of 2940±2 Ma and 2925±3 Ma (*see*

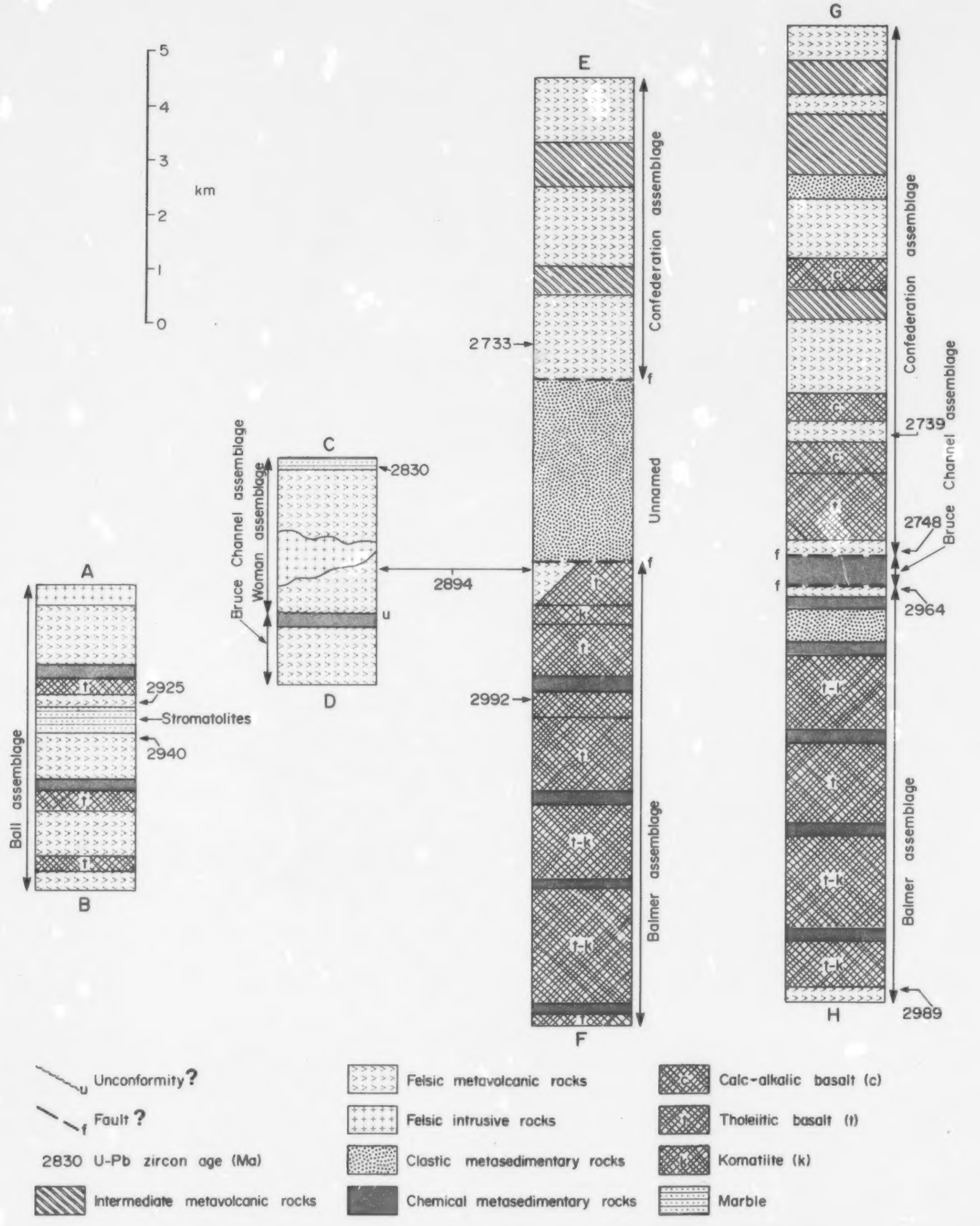

5
4
3
km
2
1
0
A
B
Ball assemblage
2925
Stromatolites
2940
C
D
Bruce Channel assemblage
Woman assemblage
2830
u
2894
E
F
Confederation assemblage
2733
f
Unnamed
Balmer assemblage
2992
G
H
Confederation assemblage
Bruce Channel assemblage
2739
2748
f
f
2964
Balmer assemblage
2989
c
t
k
t-k
Unconformity?
Fault?
2830 U-Pb zircon age (Ma)
Intermediate metavolcanic rocks
Felsic metavolcanic rocks
Felsic intrusive rocks
Clastic metasedimentary rocks
Chemical metasedimentary rocks
Calc-alkalic basalt (c)
Tholeiitic basalt (t)
Komatiite (k)
Marble

Figure 6.7. Tectono-stratigraphic sections showing typical rock units and ages of the major assemblages in the Red Lake greenstone belt. Locations of these sections are shown on Figure 6.4. (Sections *modified from* Wallace et al. 1986.)

Figure 6.4) bracket the depositional age of the shallow water stromatolites. The felsic units, above and below the marble beds, typify the main mass of calc-alkalic volcanism that dominates this assemblage and significantly contrast in age and composition with the tholeiitic basalt and komatiite flows of the older, adjacent Balmer assemblage.

The stratigraphic younging directions face outwards, away from the core of this elliptical assemblage and contrast with the northward-facing younging directions within the adjacent Balmer assemblage (*see* Figure 6.3). A fault contact is inferred as noted above. The contact with an unnamed sedimentary assemblage along the northern flank of the belt is undetermined.

Bruce Channel Assemblage (Volcanic)

Most of the Bruce Channel assemblage is poorly exposed in the eastern part of the belt (*see* Figure 6.3; Pirie and Grant 1978a, 1978b). Geophysical conductors and iron formation markers provide some control and permit a stratigraphic correlation. Much of this assemblage is composed of basaltic flows and is capped by minor units of felsic pyroclastic rocks, with a common magmatic age of 2894 Ma at 2 widely separated locations (Corfu and Wallace 1986; Corfu and Andrews 1987), and by clastic and iron formation sedimentary rocks. The eastern part of the Red Lake greenstone belt is buckle folded and the folding is accentuated by the sedimentary units in the upper part of the Bruce Channel assemblage, east of the Campbell and Arthur W. White mines (*see* Figure 6.2). The assemblage tectonically underlies the older Balmer assemblage and is in fault contact with the Confederation assemblage on its southern flank. A unit of iron formation at the top of the Bruce Channel assemblage is traced along much of its length and is the principal means of correlating the part of the assemblage on the east shore of Bruce Channel on Red Lake with the part along the eastern margin of the belt.

The Bruce Channel assemblage is also comparable in age with quartz porphyry intrusions in the Pickle Crow assemblage at Pickle Lake, but for the present, the two are treated as separate assemblages. This assemblage may correlate with volcanic rocks of the Hornby Lake greenstone belt (2901±2 Ma, D. Davis and M. Moore, Royal Ontario Museum, personal communication, 1991) in the Berens River Subprovince.

Woman Assemblage (Volcanic)

The Woman assemblage, like the Confederation assemblage, extends almost the full length of the Uchi Subprovince (*see* Figure 6.3). It occupies a substantial part of the northern half of the subprovince and occupies parts of greenstone belts with bedding that strikes northeasterly and northwesterly toward the Berens River and Sachigo subprovinces and contrasts with more easterly striking parts in the southern half of the subprovince. It may be argued that this contrast in structural grain between the north and south halves of the subprovince is attributed to a greater penetrative deformation in the south during the closing stages of the Kenoran Orogeny (*see* Sequence of Events in the Uchi Subprovince).

In the Red Lake greenstone belt, this assemblage is represented by one isolated area with 2830 million-year-old felsic volcanic rocks (Corfu and Wallace 1986) on McKenzie Island in the central part of the belt (*see* Figure 6.4). Since this felsic assemblage is surrounded by the waters of Red Lake, its relationship to the Bruce Channel assemblage and to the more dominant Balmer assemblage is not yet clear.

Confederation Assemblage (Volcanic)

The Confederation assemblage comprises the dominant volcanic assemblage in the Uchi Subprovince. This prominence is thought to reflect a comparatively greater volume of magmatism as well as greater success in preservation since the time frame encompassed by the rocks of this assemblage (2750 to 2730 Ma) also covers the period of most extensive volcanic activity preserved in the rock record across the Superior Province (*see* Map 2576, map case). The Confederation assemblage can be traced almost continuously from the Red Lake greenstone belt eastward to Pickle Lake where it probably pinches out (*see* Figure 6.3). This assemblage is characterized by a predominance of units that reflect cycles of tholeiitic to calc-alkalic volcanism with substantial products of felsic pyroclastic eruptions preserved in the uppermost units in many of the cycles. Much of this assemblage is interpreted to consist of accumulations of subaqueous shield volcanoes in an oceanic island arc setting. In some cases, the volcanic cycles are found to be tectonically repeated or stacked out of stratigraphic order. This is evident in the eastern half of the Birch–Uchi greenstone belt and in the Meen–Dempster greenstone belt as described later.

In the Red Lake greenstone belt, the Confederation assemblage is interpreted to include the Heyson calc-alkalic sequence of Pirie (1981) in the south and the Graves calc-alkalic sequence in the north, both of which Wallace et al. (1986) called Cycle III. The uppermost units of the Confederation assemblage in the south half of the belt are composed of rhyolitic pyroclastic deposits that can be correlated stratigraphically with the uppermost units of the same assemblage in the centre of the Birch–Uchi greenstone belt. Accordingly, the Red Lake part of this assemblage is inferred to share the same history.

Sedimentary Rocks

The most prominent sequences of wacke-dominated clastic sedimentary rocks are found in the northern part of the belt, especially lying between the Balmer and northern Confederation assemblages (*see* Figure 6.3). It is uncertain at present how old these sedimentary rocks are; they may be lying conformably at the top of the Balmer assemblage, although this is less likely than the possibility that they are related to the Confederation assemblage or compose an allochthonous slice of the northern English River assemblage transported northward during the Uchian phase of the Kenoran Orogeny (defined *in* Sequence of Events in the Uchi Subprovince and Tectonic Interpretation and Outstanding Problems).

Another sedimentary sequence of note appears to form the uppermost part of the Bruce Channel assemblage and is composed of wacke sandstone-mudstone interbeds and banded magnetite iron formation. It highlights the tight folding of the Balmer and Bruce Channel assemblages in the eastern part of the belt. The sedimentary sequence apparently narrows to a sulphide facies iron formation unit following the fault boundary between the Confederation and Balmer assemblages, and expands to a suite of felsic volcanic and sedimentary rocks along the east shore of Bruce Channel in Red Lake.

BIRCH–UCHI GREENSTONE BELT

The Birch–Uchi greenstone belt (*see* Figure 6.2) has been subjected to extensive volcanological and geochemical investigations, highlighted by the work of Goodwin (1967) and Thurston (1985a, 1986; Thurston and Fryer 1983). Early work, for example, by Thomson (1939b) and Bateman (1940a, 1940b), focussed on the gold deposits of the belt. Pryslak (1971a) subdivided the west half of the belt into 3 volcanic cycles; this was extended by Thurston (1985a, 1985b) to include the east half. Age determinations by Nunes and Thurston (1980) and by Corfu (Wallace et al. 1986) on the upper felsic units of each cycle (2959, 2840 and 2738 Ma) showed that these cycles were not products of a progressive volcanic episode, but rather of separate episodes of volcanism over 220 million years. Further age determinations in other parts of this subprovince (described later) and elsewhere in the Superior Province over the past decade have shown such episodicity, with greatly contrasting ages from one cycle to the next, to be the general rule. This rule implied that these cycles formed not a simple, conformable layer-cake stratigraphy (even though the cycles are superficially similar in their lithologic features) but a tectono-stratigraphy with unconformities and faults separating the cycles. However, as is still evident in the Birch–Uchi greenstone belt, the boundary characteristics between these cycles are not easily viewed and the tectonic interpretation of these boundaries is largely inferred from the age determinations and structural relationships. These cycles have now been re-interpreted in terms of tectonic assemblages and have been projected into the northern part of the belt and correlated with the assemblage patterns interpreted in the adjacent greenstone belts.

The following is a summary of the lithological and stratigraphic characteristics of these assemblages, taken mainly from Thurston (1985a) with additional observations and discussion. Stratigraphic thicknesses and dominant rock types are shown in Figure 6.8 and brief descriptions of the assemblages are given in Table 6.3.

Balmer Assemblage (Volcanic)

The Balmer assemblage in the Birch–Uchi greenstone belt is the lowermost mafic to felsic volcanic cycle (Cycle 1) identified by Pryslak (1971a, 1971b) on the western margin of the belt, adjacent to the Trout Lake batholith (*see* Figure 6.3). An age of 2959 Ma for a rhyolitic crystal lithic tuff confirms this cycle to be part of the Balmer assemblage that forms much of the Red Lake greenstone belt. This assemblage in the Birch–Uchi greenstone belt is up to 4800 m thick and has been subdivided by Thurston (1985a, 1985b) into 4 formations (A to D) of basaltic flows, wacke sandstone, intermediate to felsic pyroclastic rocks and chemical sediments, summarized in Table 6.3.

Woman Assemblage (Volcanic)

In the Birch–Uchi greenstone belt, the upper felsic volcanic unit of the Woman assemblage has been dated at about 2840 Ma (F. Corfu, unpublished data referred to by Wallace et al. 1986) and corresponds in age with a thin unit of fine-grained (distal?) felsic tuff in the northern Meen–Dempster greenstone belt (*see* Figure 6.3). However, the latter is overlain by another basalt-dacite cycle, dated at about 2825 Ma; it is thus apparent that the boundary with the overlying Confederation assemblage, which extends along much of the length of the subprovince, does not exclusively follow the top of the 2840 million-year-old cycle. This boundary with the Confederation assemblage may be a paraconformity in some places and a fault in others. It is possible that there was no deposition of the 2825 million-year-old cycle in the Birch–Uchi greenstone belt; alternatively, this cycle may have been removed or overlain by an allochthonous or parautochthonous Confederation assemblage transported from the south. Along the length of Woman Lake, which coincides with the boundary between the Woman and Confederation assemblages in the Birch–Uchi greenstone belt, Fyon and O'Donnell (1986) identified numerous high strain zones that may reflect later reactivation of a probable fault along this boundary.

The Woman assemblage, like the other younger assemblages, is composed of mafic to felsic volcanic cycles typically with chemical and tuffaceous metasedimentary rocks at the top. These cycles comprise lower units of tholeiitic basalts and minor basaltic andesites; and upper units of calc-alkalic pyroclastic deposits and local flows. The Woman assemblage in the west half of the Birch–Uchi greenstone belt has been subdivided by Thurston (1985a) into 4 formations—E, G, H and J, comprising mafic and intermediate flows, and intermediate to felsic pyroclastic deposits capped by chemical sedimentary rocks. They are summarized in Table 6.3. Thurston (1985a) originally interpreted the eastern half of the Birch–Uchi greenstone belt in terms of the 3-cycle subdivision developed originally in the west half, since Thurston saw comparable lithologic features on both sides of the belt. However, Noble (1989) demonstrated by dating 2 felsic units that the eastern half of the belt is a thickly stacked sequence of volcanic cycles, comparable in age with Thurston's Cycle III (Confederation assemblage). Consequently, some of the formations or parts of their descriptions by Thurston have been deleted or revised for Table 6.3 so that they apply only to the stratigraphy in the west half of the belt.

The Woman assemblage features a well-preserved, thickly bedded sequence of rhyolitic tuff deposits described by Thurston (1980a) to compare in its internal fabric construction with subaerial ignimbrites produced by pyroclastic flow as described by Sparks et al. (1973). These bedding and fabric features include poorly developed stratification, reverse size grading with pumiceous fragments concentrated at the tops of flow units, and abundant flattened pumiceous clasts forming fiamme accompanied by devitrified glass shards in welded to partly welded zones. This 300 m thick tuff is interpreted by Thurston (1980a, 1980b) to be the product of a Plinian eruption, and deposited subaerially or in shallow water. This unit is overlain by shallow water carbonate on Woman Lake; a limited part of the carbonate unit is noted by Hofmann et al. (1985) to contain small mounded structures that can be attributed to microbial activity typical of stromatolites.

Confederation Assemblage (Volcanic)

The Confederation assemblage (*see* Figure 6.3) had been interpreted by Thurston (1985a) to be centred on a broad synclinorium in the Birch–Uchi greenstone belt with an anticline on the east side of the belt. The interpretation of the synclinorium was based in part on the presence of opposing stratigraphic top directions along its length. However, ages are not symmetrically distributed across this synclinorium as was once assumed from U-Pb ages determined on its west side (Nunes and Thurston 1980). Noble et al. (1989) have now shown that the eastern half of the belt is composed of

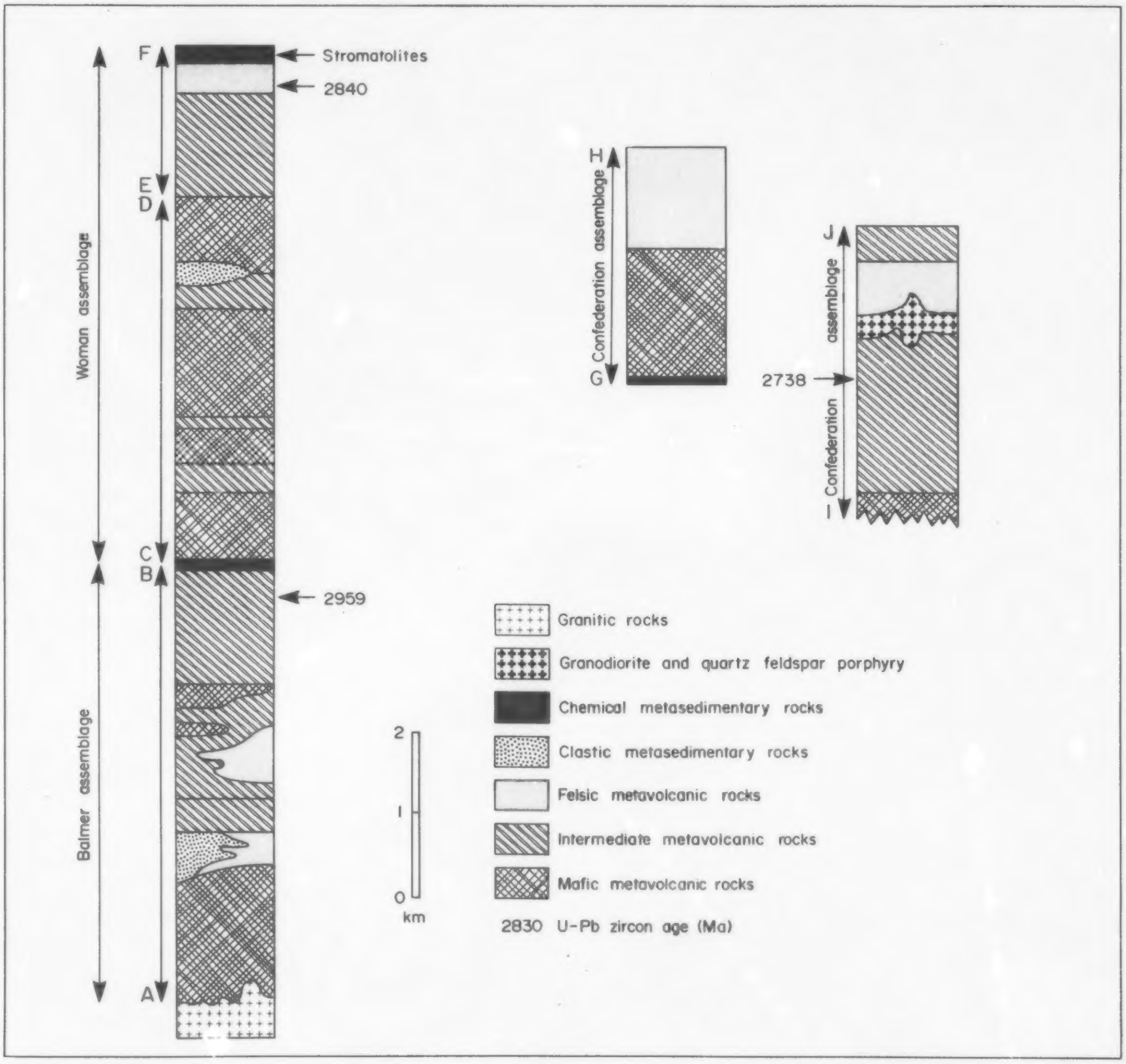

F
Stromatolites
2840
E
D
Woman assemblage
C
B
2959
Balmer assemblage
A
H
Confederation assemblage
G
J
assemblage
2738
Confederation
I
2
1
0
km
Granitic rocks
Granodiorite and quartz feldspar porphyry
Chemical metasedimentary rocks
Clastic metasedimentary rocks
Felsic metavolcanic rocks
Intermediate metavolcanic rocks
Mafic metavolcanic rocks
2830 U-Pb zircon age (Ma)

Figure 6.8. Tectono-stratigraphic sections showing typical rock units and ages of the assemblages in the Birch–Uchi greenstone belt. Locations of these sections are shown on Figure 6.5 (*modified from* Thurston 1986).

either a complex interleaving of several rapidly deposited volcanic cycles or a tectonic repetition of a single cycle in the Confederation assemblage; an assessment of the true nature of this tectono-stratigraphic framework remains to be evaluated since all 3 U-Pb age determinations in this region of the Confederation assemblage lie within the errors of precision (*see* Table 6.3 and Figure 6.5).

The rock units of the Confederation assemblage in this belt have been summarized in several geological reports by Thurston (1985a, 1985b), Good (1988) and Beakhouse (1989). This assemblage features several mafic to felsic volcanic cycles. The geochemical studies of Thurston and Fryer (1983) showed that across the 3 volcanic assemblages in this belt, there are distinct differences in composition of the cycles and differences in the petrological processes involved with their formation. Indeed, Thurston (1986) encouraged the view of Anhaeusser (1971) that there are several scales of cyclicity in the evolution of this and other greenstone belts. Thurston (1986) gave an example of a major cycle several thousand metres thick in the centre of the belt where a fault-bounded felsic pyroclastic sequence was built upon a shield volcano constructed largely of mafic to intermediate pillowed flows; the pyroclastic sequence is marked by the prominence of ash-flow deposits and vent-related flows, and is interpreted to be a caldera complex that evolved in a subaqueous environment, culminating in a Plinian eruption followed by caldera collapse (Thurston 1986). The chronology of events compares favourably with the resurgent cauldron cycle of Smith and Bailey (1968). The cyclicity of volcanism and the ensuing products are typical of island arc volcanism; Cycle III (Woman and Confederation assemblages) has been interpreted to be comparable with such an oceanic arc environment by Thurston and Chivers (1990).

The subaqueous pyroclastic rocks of the Confederation assemblage are notable because of their excellent preservation and the lack of reworking and resedimentation after eruption. Stix and Gorton (1989) described sections of a monolithic sequence of felsic pyroclastic rocks in the eastern half of this assemblage on Lost Bay of Confederation Lake. The pyroclastic sequence examined is composed of numerous flow deposits, each less than 3 m thick and displaying graded bedding and stratification that imply the pyroclastic ash was mixed with water either in the eruption column or at the time of deposition. The authors observed no evidence of welding, confirming the substantial mixing of the flows with water, and concluded that, unlike the subaerial welded Woman Lake tuff beds documented by Thurston (1980a), the sequence on Lost Bay is composed of subaqueous pyroclastic debris flows.

In the northern part of the belt, at the junction between Springpole Lake and Birch Lake (*see* Figure 6.2; Good 1988), there is a locally developed complex, only about 5 km long, of trachyte flows, syenite dikes, a fluorite-bearing carbonatite intrusion and associated fenite breccia and other intrusive breccia. This was investigated recently by Barron (in press) and Barron et al. (1989), and attracted some gold exploration interest, although it was described many years earlier by Harding (1936a). The trachytic volcanic rocks include potassium-feldspar-phyric flows, coarse vent facies breccia, lapilli tuff and fine tuff at the base of an unconformity between the underlying Confederation assemblage rocks and younger sedimentary rocks consisting of polymictic conglomerate and sandstone. Similar alkalic magmatism is observed to be associated with late cratonization at the terminating stages of the Kenoran Orogeny in the Abitibi Subprovince near Kirkland Lake (the Timiskaming assemblage, *see* Jackson and Fyon, this volume) and in the Shebandowan greenstone belt near Thunder Bay (Shegelski 1980). However, Barron (in press) reported a U-Pb age of 2734^{+24}_{-19} Ma for this alkalic suite which compares closely with the age of Confederation assemblage volcanism. Its spatial and temporal relation to the Confederation assemblage implies that the alkalic rocks developed within a volcanic arc and bear no relation to the late-tectonic setting attributed to the Timiskaming Group (assemblage) of the Abitibi greenstone belt (Corfu et al. 1991).

The western boundary of the Confederation assemblage with the Woman assemblage in the central part of the belt may be a fault owing to the concentration of shear zones striking obliquely across this boundary (Fyon and O'Donnell 1986). The tholeiitic basalt base of the Confederation assemblage overlies the upper felsic volcanic units and a local marble unit at the top of the Woman assemblage. To the east, the Confederation assemblage is tectonically interleaved with an assemblage of English River-type wacke turbidites as described later.

Whereas the western half of the Confederation assemblage in this belt is composed of a single cycle about 7000 m thick, the eastern half of the assemblage is marked by multiple mafic to felsic cycles, each generally less than 1500 m thick. The stratigraphic patterns and folds in the eastern half, plus the small range of the age determinations (Nunes and Thurston 1980; Noble 1989) lead one to infer that the eastern half of this assemblage, and possibly its extension south of Red Lake, has been subjected to considerable stacking and crustal thickening through faulting followed by folding.

Sedimentary Rocks

The easternmost part of the Birch–Uchi greenstone belt is dominated by thickly interleaved wacke-mudstone sequences and fault panels of basaltic rock. Descriptions of the northern part of this unnamed assemblage are given by Good (1988) and the southwestern part by Bowen (1989) although much of the assemblage awaits detailed investigation. We postulate that the interleaved sedimentary and basaltic rocks have been allochthonously or parautochthonously transported northward during the Kenoran Orogeny to form a northward-protruding finger of tectonically layered rock with its eastern part incised by external granitoid plutons; in this scenario, the sedimentary rocks were largely derived from the English River sedimentary prism south of the Uchi Subprovince.

MEEN–DEMPSTER AND PICKLE LAKE GREENSTONE BELTS

The Meen–Dempster and Pickle Lake greenstone belts (*see* Figure 6.2) form the northern half of an extensively intruded supracrustal block in the central part of Uchi Subprovince; this block is virtually separated by plutons from the Birch–Uchi and Miminiska–Fort Hope greenstone belts. The Meen–Dempster greenstone belt was first mapped by Harding (1936b) after a Red Lake prospector advised the Ontario government that he discovered volcanic rocks north of the much travelled Lake St. Joseph. In 1972, the belt was included as part of a helicopter-supported reconnaissance mapping program (Sage and Breaks 1982) and more recently remapped by Stott and LaRocque (1983a, 1983b), Stott and Wilson (1986a, 1986b) and Stott et al. (1989a). The Pickle Lake greenstone belt received attention principally in the vicinity of the Central Patricia and Pickle Crow gold mines and nearby prospects where bedrock exposures are also more abundant. Hence, mapping by Thomson (1939a), Ferguson (1966), Pye (1976) and others, summarized by Sage and Breaks (1982), focussed on the areas of gold mineralization. The Meen–Dempster greenstone belt forms a stratigraphic continuum with the Pickle Lake greenstone belt, although there are differences between the 2 and their stratigraphic relationships have not been fully resolved yet. The volcanic rocks are dominantly tholeiitic to calc-alkalic with scarce komatiite, principally in the Pickle Lake greenstone belt.

Assemblage correlations as shown in Figure 6.3 are permitted by: an analysis of the stratigraphy from previously published maps and reports of this belt summarized by Sage and Breaks (1982) and Stott et al. (1986a, 1986b, 1987a, 1987b, 1989a, 1989b); aeromagnetic and electromagnetic patterns (OGS 1986); and selected isotopic age determinations (Corfu and Stott 1989 and other unpublished determinations by F. Corfu, Royal Ontario Museum). The Meen–Dempster and Pickle Lake greenstone belts have been subdivided into 4 volcanic assemblages of diverse ages, plus a much younger sedimentary assemblage, and an area of volcanic rocks (in the northern Pickle Lake greenstone belt) that may be an assemblage, but remains unassigned until its age is clarified and its potential assignment to a named assemblage has been tested. The tectonic boundaries between assemblages in the Meen–Dempster greenstone belt are chosen to coincide with the tops of felsic volcanic units that cap basalt-dacite-rhyolite cycles; the felsic volcanic units have isotopic ages that contrast markedly with the ages of similar units in adjacent cycles. The thick sequences of tholeiitic basalt flows are assumed to form an original base upon which the volcanic edifices were constructed, comprising mainly extrusive and pyroclastic rocks of intermediate to felsic composition.

Although there is a general southward younging of assemblages towards the boundary with the English River Subprovince, the Meen–Dempster and Pickle Lake greenstone belts show a more complex arrangement (*see* Figure 6.3). For example in the Bamaji Lake area, a part of the 2840 to 2806 million-year-old Woman assemblage is enveloped by the younger Confederation assemblage and separated from the main semicontinuous tract of this assemblage farther north. Also, in the 40 km long Lang Lake greenstone belt, just north of Meen Lake (Figure 6.9), an age of about 2749 Ma for a dacitic tuff reveals that some younger supracrustal rocks lie north of most of the older Woman assemblage in the subprovince. If a southward younging of assemblages is the fundamental characteristic of this subprovince, then these exceptions need explanation. Discussion on this topic is found in the section entitled Tectonic Interpretation and Outstanding Problems. The following are highlights of some of the features and inferences arising from reconnaissance mapping of these 2 belts. The essential characteristics of the assemblages are found in Table 6.4.

Older(?) Volcanic Rocks of the Pickle Lake Area

The northern Pickle Lake area (*see* Figure 6.3) has not been assigned to an assemblage because it may in future be shown to be correlative with similar 2900 to 3000 million-year-old supracrustal rocks of the North Caribou terrane in the Sachigo Subprovince. This correlation is postulated on the basis of aeromagnetic anomalies that may reflect supracrustal extensions of the belt, which trend northeastward along with the structural grain of enveloping tonalite gneisses, towards the southeasternmost tail of the North Caribou greenstone belt. This unnamed northern half of the Pickle Lake area is dominantly composed of massive and pillowed tholeiitic basalt flows, accompanied by magnetite-banded iron formation interbeds, sill-like mafic to ultramafic intrusive sheets and very limited exposures of dacitic to rhyolitic tuff beds (Stott et al. 1989a, 1989b). Most units of this area are moderately deformed with only local zones of pervasive hydrothermal alteration. Several long, linear aeromagnetic anomalies and electromagnetically conductive graphitic schist beds and other marker beds continue for many kilometres in this area; these together with other correlated stratigraphic units indicate that the northern Pickle Lake region is typically composed of tabular volcanic and minor sedimentary strata that extend along almost the full length of the northern half of Pickle Lake greenstone belt. However, there is no clear determination of the general sense of stratigraphic younging in these rocks. There is also no evidence of the age of these rocks; they are interpreted here to be potentially correlative with the oldest strata in the Uchi or Sachigo subprovinces, such as the Balmer and McGruer assemblages (*see* Thurston, Osmani et al., this volume).

The predominance of subaqueous basaltic flows with minor vesiculation, the long, undisrupted iron formation interflow units, the very limited presence of clastic sedimentary detritus, and lack of a central calc-alkalic volcanic edifice are characteristics of an oceanic mafic plain environment (cf., Thurston and Chivers 1990), typically found in modern back arc basins for example.

This area is bounded to the south by the Pickle Crow assemblage (*see* Figure 6.3). The boundary between these 2 domains is determined from the juxtaposition of their different structural trends, both stratigraphic and geophysical; the northeastward strike of the northern Pickle Lake basalts and iron formation marker beds is at a 15° to 30° angle counterclockwise to the more easterly striking Pickle Crow units (Figures 6.10a and 6.10b). Based on this angular relationship, one could infer that the boundary is a tectonic juxtaposition of separate assemblages and if validated, this approach could be helpful in identifying other cryptic tectonic boundaries in the greenstone belts. Another example is described later from the northern Miminiska–Fort Hope greenstone belt.

Pickle Crow Assemblage (Volcanic)

The Pickle Crow assemblage (*see* Figure 6.3) consists of a folded sequence of subaqueous basaltic flows, dacitic to rhyolitic pyroclastic flows, synvolcanic quartz porphyry intrusions, and magnetite-banded iron formation plus thin layers of graphitic schist. There is limited evidence of sedimentary carbonate layers, spatially associated with the iron formation, that may have developed on a shallow-water platform (D. Mullen, Umex Incorporated, personal communication, 1988). A suite of lenticular, quartz porphyry sills intruded the basaltic base of this assemblage and one sill has been dated at about 2892 Ma (F. Corfu, unpublished data), an age comparable to that of rhyolitic tuff in the Bruce Channel assemblage of Red Lake (*see* Figure 6.4) and to that of the Pembina Tonalite Gneiss in the Lake St. Joseph batholith. The Pickle Crow assemblage is host to the Central Patricia and Pickle Crow gold deposits and is the locus of extensive hydrothermal alteration, folding and zones of ductile deformation trending subparallel to the strata. The oxide facies banded iron formation units provide markers to trace the folded stratigraphy in this assemblage.

Woman Assemblage (Volcanic)

The Meen–Dempster greenstone belt (*see* Figure 6.2) is dominated by the Woman assemblage and the overlying Confederation assemblage (*see* Figure 6.3). Both assemblages, striking eastward, face (young) stratigraphically southward, but are overturned and dip steeply to the north; the facing direction of the Woman assemblage in this belt is consistent with that determined in the Birch–Uchi and Pickle Lake greenstone belts. The lower half of the Woman assemblage is characterized by massive to pillowed tholeiitic basalt flows, associated gabbroic sills and intervening thin beds of banded, quartz-magnetite iron formation and mudstone. The iron formation beds can be traced discontinuously as a marker unit in the basalt for 50 km along strike (Stott and LaRocque 1983a, 1983b; Stott and Wilson 1986a); the iron formation unit overlies, within 1000 m, a 1 m wide felsic tuff bed isotopically dated at 2841^{+5}_{-2} Ma, and underlies by about 1000 m, the dacitic to rhyolitic pyroclastic rocks at the top of this assemblage, dated at 2825±2 Ma (F. Corfu, unpublished data). The iron formation may mark a significant hiatus in volcanic activity of this assemblage; such time breaks in this and other greenstone belts may also prove, with more detailed investigations, to

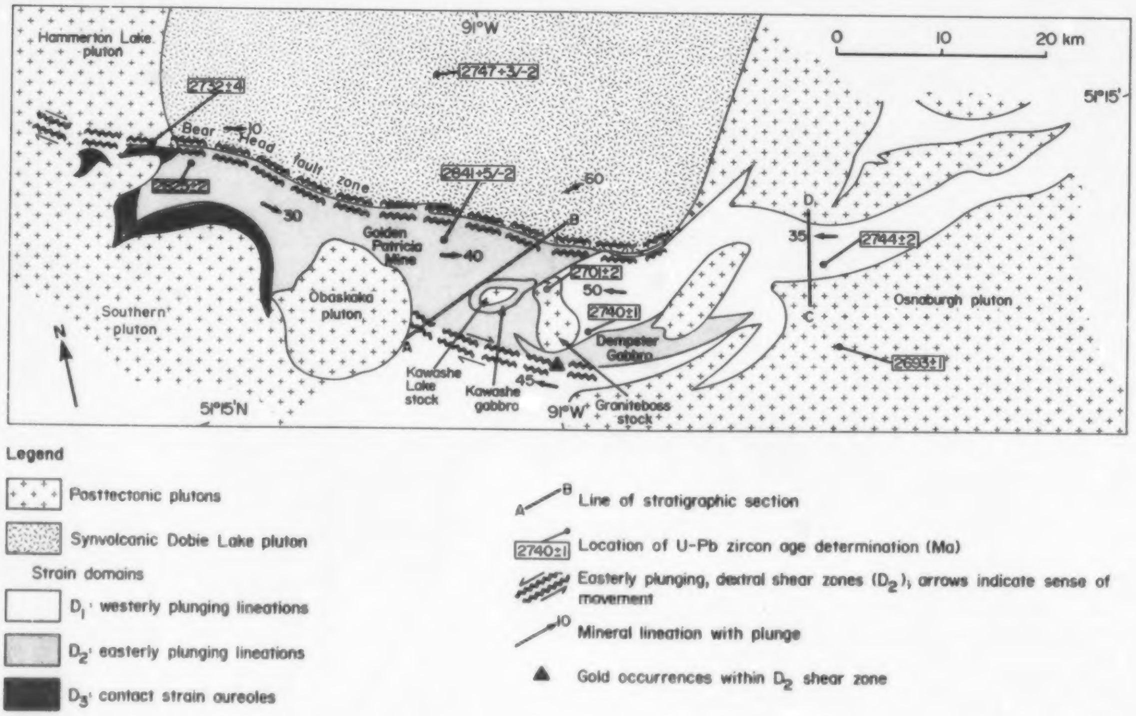

Hammerton Lake pluton
91°W
2747+3/-2
2732±4
Bear Head fault zone
10
2825±2
2841+5/-2
60
30
B
Golden Patricia Mine
40
2701±2
50
2740±1
Dempster Gabbro
Obaskaka pluton
Southern pluton
N
A
Kawashe Lake stock
Kawashe gabbro
45
51°15'N
91°W
Graniteboss stock
0
10
20 km
51°15'
D
35
2744±2
C
Osnaburgh pluton
2693±1
Legend
Posttectonic plutons
Synvolcanic Dobie Lake pluton
Strain domains
D1: westerly plunging lineations
D2: easterly plunging lineations
D3: contact strain aureoles
A B Line of stratigraphic section
2740±1 Location of U-Pb zircon age determination (Ma)
Easterly plunging, dextral shear zones (D2), arrows indicate sense of movement
10 Mineral lineation with plunge
Gold occurrences within D2 shear zone

Figure 6.9. The Meen–Dempster greenstone belt, showing tectonic strain domains, locations of age determinations and the locations of stratigraphic sections shown in Figure 6.11. The domains of strain are each characterized by consistent orientations of stretching lineations, as shown schematically.

be fundamental paraconformities between volcanic sequences that originated from contrasting depositional environments. This cannot be demonstrated in the Meen–Dempster greenstone belt; instead, the basaltic flows and iron formation are considered typical of oceanic environments, including back arcs or basal platforms upon which island arcs grew.

The uppermost pyroclastic sequence of the Woman assemblage is up to 1.4 km wide on Meen Lake, where proximal facies deposits are concentrated, and thins along strike to about 200 m wide of more distal fine-grained dacitic tuff. Rhyolite is accompanied on the upper flanks of 2 eroded, volcanic edifices by feldspathic arenite, scarce marble and ferruginous chert beds. The thicker stratigraphic parts of this upper calc-alkalic sequence coincide with the coarser, proximal deposits on Meen Lake (Stott and LaRocque 1983a) and 24 km farther east where a smaller volcanic centre (Stott and Wilson 1986a) is inferred south of Muskegsagagen Lake.

In the Pickle Lake greenstone belt, this assemblage is represented most clearly along the southeastern flank of the belt where a 2200 m wide sequence of basaltic flows and thinner dacitic to rhyolitic ash flows were deposited. Quartz-phyric, felsic pyroclastic rocks locally contain volcanic massive sulphide lenses, which are dominated by massive pyrrhotite and overlain by thin chert beds; the stratigraphic relationships indicate a primary, southward-younging direction. A quartz-phyric tuff has provided a U-Pb zircon age of about 2836 Ma (F. Corfu, unpublished data). It is suggested here that a possible stratigraphic correlation exists between the felsic volcanic rocks of the Woman assemblage in the Pickle Lake greenstone belt and a coarser, thicker suite of dacitic pyroclastic rocks and a quartz porphyry intrusion, inferred to lie close to a vent, in an unnamed assemblage in the northern part of the Miminiska–Fort Hope greenstone belt.

The southern part of the Meen–Dempster greenstone belt, in the Bamaji Lake area (*see* Figure 6.2), contains a progressively southward-younging part of the Woman assemblage. Two dacitic tuff units have been dated by Scharer (1989) at about 2816 and 2805 Ma. This southern part of the Meen–Dempster greenstone belt is partly underlain by a large subvolcanic, molybdenite-bearing trondhjemite pluton on North Bamaji Lake, with an age by Scharer (1989) of about 2806^{+18}_{-14} Ma.

In summary, the Woman assemblage appears to be comparable to island arc chains, built upon an oceanic basaltic platform with several calc-alkalic edifices formed over 40 million years; the edifices locally breached the water surface, eroded to form sandstone beds, and provided shallow water conditions on their flanks for the formation of graphitic mudstone, iron formation and carbonate beds.

DISCUSSION

The youngest Woman assemblage age of 2805 Ma (Scharer 1989) compares with the age of 1 tonalite phase in the Trout Lake batholith (2806 Ma, Noble et al. 1989), which straddles and intrudes the contact between the Balmer and Woman assemblages in the Birch–Uchi greenstone belt. It has been postulated by Thurston, Osmani et al. (this volume) that 2805 Ma marks not only the termination of Woman assemblage volcanic activity, but also the beginning of collision between the Woman assemblage and the older Balmer assemblage, which forms the southernmost flank of the North Caribou terrane. Relevant to this scenario is the presence of the 2781 million-year-old Bamaji assemblage (Scharer 1989), described below, which forms along the southernmost exposure of the Woman assemblage in the Uchi Subprovince and could be interpreted as an intracratonic pull-apart basin. It does not, however, appear to mark the southernmost extent of 2800 million-year-old continental crust, since the 2887 million-year-old Pembina Tonalite Gneiss lies even farther south in the Lake St. Joseph batholith. The position taken in this chapter, however, is that the Woman assemblage developed on the margin of a proto-continent, which included the Balmer assemblage, and that the Woman assemblage is either autochthonous or a parautochthonous slice transported northward. The evidence for its continental arc setting is the presence of a 2838 million-year-old tonalite body in the Trout Lake batholith (Noble 1989), synchronous with volcanism of the Woman assemblage which intrudes the Balmer assemblage, underlying the Woman assemblage in the Birch–Uchi greenstone belt. It is argued that the juxtaposition of the Balmer and Woman assemblages occurred at the time of *circa* 2740 million-year-old Woman assemblage magmatism.

Bamaji Assemblage (Volcanic and Sedimentary)

The Bamaji assemblage is a thin wedge of metavolcanic and metasedimentary rocks lying between the Woman assemblage at Bamaji Lake and the Bamaji–Blackstone batholith to the south (*see* Figures 6.2 and 6.3). A felsic tuff in this assemblage, dated at 2781^{+4}_{-3} Ma (Scharer 1989), is significantly younger than the underlying Woman assemblage rocks. Its relationship to the Confederation assemblage toward Lake St. Joseph is not known.

The Bamaji assemblage, typically about 700 m thick, is composed of a basal conglomerate unit, potassium-rich (3.5%) dacitic pyroclastic rocks and thinly bedded wacke. The conglomerate is draped along the flanks of the trondhjemitic North Bamaji pluton (*see* Figure 6.2) and consists of detritus derived from the exhumation of the pluton.

The presence of this proximal conglomerate implies that the Bamaji assemblage was built on an exposed continental crust. The unusually high potassium content of the pyroclastic rocks is consistent with the extrusion of alkalic to calc-alkalic volcanic material in a continental rift. This assemblage thus appears to be a late orogenic suite, unconformably overlying the 2805 million-year-old Woman volcanism and marking the termination of a period of orogenesis from about the end of 2805 Ma volcanism to 2781 Ma.

Confederation Assemblage (Volcanic)

The Confederation assemblage in the Meen–Dempster greenstone belt (*see* Figure 6.3) is composed of 2 mafic to

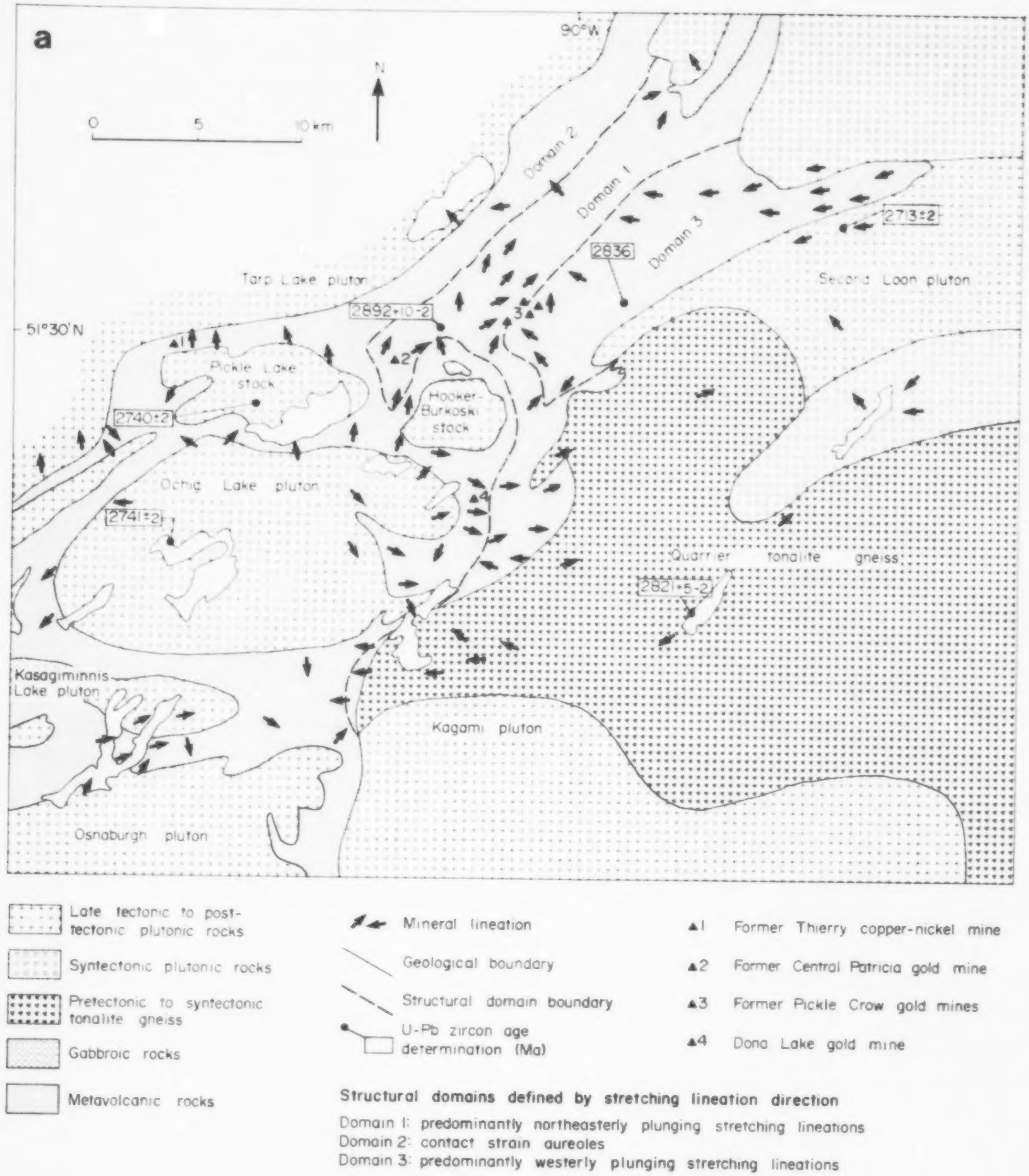

a
90°W
N
0
5
10 km
Domain 2
Domain 1
Domain 3
2836
2713±2
Tarp Lake pluton
Second Loon pluton
2892+10-2
3
51°30'N
1
Pickle Lake stock
2
Hooker-Burkoski stock
2740±2
Ochig Lake pluton
4
2741±2
Quarrier tonalite gneiss
2821+5-2
Kasagiminnis Lake pluton
Kagami pluton
Osnaburgh pluton
Late tectonic to post-tectonic plutonic rocks
Syntectonic plutonic rocks
Pretectonic to syntectonic tonalite gneiss
Gabbroic rocks
Metavolcanic rocks
Mineral lineation
Geological boundary
Structural domain boundary
U-Pb zircon age determination (Ma)
▲1 Former Thierry copper-nickel mine
▲2 Former Central Patricia gold mine
▲3 Former Pickle Crow gold mines
▲4 Dona Lake gold mine
Structural domains defined by stretching lineation direction
Domain 1: predominantly northeasterly plunging stretching lineations
Domain 2: contact strain aureoles
Domain 3: predominantly westerly plunging stretching lineations

Figure 6.10. a) The Pickle Lake greenstone belt, showing tectonic strain domains, locations of age determinations and locations of former and active mines. The strain domains are recognized as areas of characteristic orientations of stretching lineations, shown here schematically. The area containing the earliest recognized strain fabrics is limited mainly to the narrow domain 1 in the northern part of the belt. Domain 2 is dominated by the strain aureole around the Ochig Lake pluton (*see* Figure 6.15). Domain 3 strain fabric is attributed to the emplacement of the Second Loon pluton. The Central Patricia–Pickle Crow gold deposits principally lie within domain 1 and the folded veins hosting the mineralization plunge steeply to the northeast. Part of the Pickle Crow deposits lie within strain domain 3 and show folded veins that plunge to the northwest. The complexity of strain domains highlights the difficulty in defining areas that still preserve an early record of deformation in volcanic belts. *See* text for further discussion. **b)** Magnetite iron formation and electromagnetic conductors in the north half of the Pickle Lake greenstone belt are at an angle to the conductors in the Pickle Crow assemblage, thereby identifying the assemblage boundary as a fault.

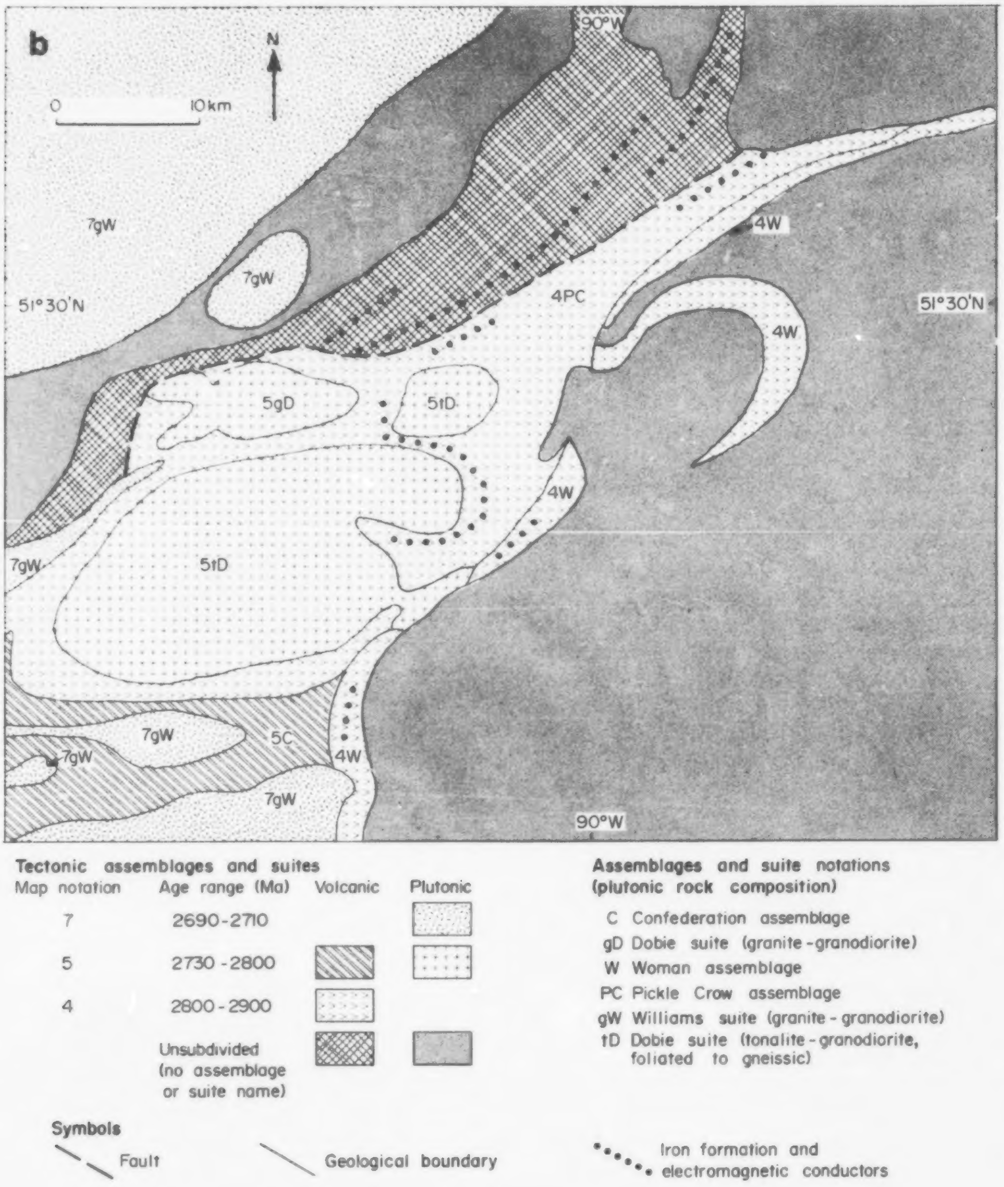

b
N
0
10km
90°W
7gW
7gW
51°30'N
4W
4PC
51°30'N
4W
5gD
5tD
4W
7gW
5tD
7gW
7gW
5C
4W
7gW
90°W
Tectonic assemblages and suites
Map notation
Age range (Ma)
Volcanic
Plutonic
7
2690-2710
5
2730-2800
4
2800-2900
Unsubdivided (no assemblage or suite name)
Assemblages and suite notations (plutonic rock composition)
C Confederation assemblage
gD Dobie suite (granite-granodiorite)
W Woman assemblage
PC Pickle Crow assemblage
gW Williams suite (granite-granodiorite)
tD Dobie suite (tonalite-granodiorite, foliated to gneissic)
Symbols
Fault
Geological boundary
Iron formation and electromagnetic conductors

Figure 6.10. a) The Pickle Lake greenstone belt, showing tectonic strain domains, locations of age determinations and locations of former and active mines. The strain domains are recognized as areas of characteristic orientations of stretching lineations, shown here schematically. The area containing the earliest recognized strain fabrics is limited mainly to the narrow domain 1 in the northern part of the belt. Domain 2 is dominated by the strain aureole around the Ochig Lake pluton (*see* Figure 6.15). Domain 3 strain fabric is attributed to the emplacement of the Second Loon pluton. The Central Patricia–Pickle Crow gold deposits principally lie within domain 1 and the folded veins hosting the mineralization plunge steeply to the northeast. Part of the Pickle Crow deposits lie within strain domain 3 and show folded veins that plunge to the northwest. The complexity of strain domains highlights the difficulty in defining areas that still preserve an early record of deformation in volcanic belts. *See* text for further discussion. **b)** Magnetite iron formation and electromagnetic conductors in the north half of the Pickle Lake greenstone belt are at an angle to the conductors in the Pickle Crow assemblage, thereby identifying the assemblage boundary as a fault.

intermediate volcanic cycles overlain by a basaltic unit that forms either a mafic volcanic cap to the second cycle or the lower mafic part of a dismembered third cycle along the southern flank of the belt. Stratigraphic tops from mafic pillow shapes in each of the volcanic cycles face southward; this part of the assemblage is inferred to lie on strike with the continuation of the easterly facing west half of the assemblage in the Birch–Uchi greenstone belt.

The Confederation assemblage in the Meen–Dempster greenstone belt is in apparent fault contact on its north margin with the Woman assemblage. The uppermost part of the Woman assemblage comprises well preserved beds of tuff and metasediment as described earlier; this pinches out to the east where (*see* Figure 6.3) the Woman assemblage strata appear to be obliquely transected by the overlying Confederation assemblage near the Graniteboss stock (*see* Figure 6.6). The southern boundary of the Confederation assemblage is an angular unconformity marked by a younger sequence of folded wacke turbidite deposits, iron formation beds and proximal submarine fan deposits of the Billett assemblage. The northern boundary of these metasedimentary rocks obliquely transects the 2 Confederation assemblage cycles.

Dacitic ash flows dominate the upper units of the 2 cycles in this belt. They show similar ages, 2740±1 and 2744±2 Ma (F. Corfu, unpublished data), with the older of the 2 determinations lying stratigraphically above the younger. One could infer from these ages that either the 2 cycles are derived from a single volcanic vent that has been tectonically repeated by imbricate stacking or they represent 2 rapidly deposited sequences. It is noteworthy that Noble (1989; Noble et al. 1989) has demonstrated that the eastern half of the Birch–Uchi greenstone belt is composed of multiple cycles of the Confederation assemblage that may comprise tectonically stacked units. This assembly of a greenstone belt by tectonic stacking is also evident on Lake St. Joseph (*see* below).

The Confederation assemblage in this belt possesses whole rock geochemistry similar to the Woman assemblage (Stott, unpublished data; Wallace 1985). Each assemblage is principally characterized by a base of tholeiitic basalt flows overlain by calc-alkalic dacite to rhyolite pyroclastic deposits. Komatiitic rocks are rare. The Confederation assemblage stratigraphy is indicated in Figure 6.11. It is treated for convenience in terms of 2 cycles. The base of Cycle 1 comprises 1400 to 2300 m of mainly pillowed tholeiitic basalt flows; the flows are rich with iron carbonate-filled vesicles. The upper part of the cycle is dominated by 1000 to 2000 m of massive, thick beds of dacitic tuff (probably ash-flow deposits). This calc-alkalic sequence thickens and coarsens toward the eastern part of the belt. Cycle 2 is similar, consisting of 1000 to 1400 m thick, pillowed basaltic flows overlain by a banded magnetite iron formation unit and 1200 to 2100 m of dacitic ash-flow deposits that locally change to andesitic composition upward in the section. The Confederation assemblage in this belt appears to represent a subaqueous basaltic basin upon which was draped the medial to distal ash deposits from a calc-alkalic volcanic edifice, the location of which is not known.

Billett Assemblage (Sedimentary)

A late orogenic sequence of clastic sedimentary rocks and interbeds of banded magnetite iron formation transects the Meen–Dempster greenstone belt volcanic assemblages with angular unconformity on a regional scale. The Billett assemblage (*see* Figure 6.3) principally contains turbiditic wacke beds, but local proximal conglomerate and graded sandstone interbeds (Figure 6.12) along its northern margin may be deposits in a submarine fan. The presence of a tonalite cobble dated at 2748^{+5}_{-4} Ma, which compares with the Dobie Lake pluton to the north, implies that the source of the clastic debris is from exposed continental crust close by. Inclusions of this assemblage can be traced across part of the Lake St. Joseph batholith; correlation with other synorogenic sedimentary deposits (e.g., in the Lake St. Joseph and Birch–Uchi greenstone belts) is speculative.

LAKE ST. JOSEPH GREENSTONE BELT

Confederation Assemblage (Volcanic)

In the western part of Lake St. Joseph (*see* Figure 6.2), there are 4 volcanic cycles (*see* Figure 6.6). These were originally mapped and described by Clifford (1969; Clifford and McNutt 1971) who recognized sequences of Upper and Lower Volcanics and an overlying sequence of Upper Clastic Rocks. Berger (1981) recognized additional top determinations which permitted subdividing this region into 3 volcanic cycles. An additional, younger volcanic Cycle 4 was traced by Stott et al. (1987a, 1987b) along the length of Lake St. Joseph and it can also be partly inferred from the original mapping by Clifford (1969). The first 3 mafic to felsic volcanic cycles define an eastward-facing, steeply plunging anticline (*see* Figure 6.6) that is overturned and verging steeply southward. When ages were determined by the second author (F. Corfu, 1989) for the upper felsic units of cycles 1 and 3, it was discovered that they compare closely (2733^{+3}_{-2} Ma and 2729.5±1 Ma, respectively) and yet differ significantly from the age of an upper felsic unit of the intervening Cycle 2 (2722±2 Ma), part of the St. Joseph assemblage (*see* Figure 6.3). The implication is that the volcanic cycles had been tectonically shuffled out of sequence as described in the section on structural geology.

Cycles 1 and 3 are comparable in age and geochemical signatures (Stott, unpublished data) and confirm the existence of a lower volcanic sequence as originally postulated from work done on the geochemistry and field relations by Clifford and McNutt (1971). This part of the Confederation assemblage is slightly younger than the eastern half of the Birch–Uchi greenstone belt, yet similar in age to the Bidou Lake Subgroup of the Rice Lake greenstone belt in Manitoba (Turek et al. 1989). Cycle 1 of this belt is physically continuous with strata along the southern flank of the Birch–Uchi greenstone belt.

The volcanic paleotopography appears to have been partly preserved in cross section; the edifice of the shield volcano, now tilted on its side and composed of Cycle 1 and a disconformable Cycle 2 representing reactivation of the volcano, is centred with its proximal and vent facies in the

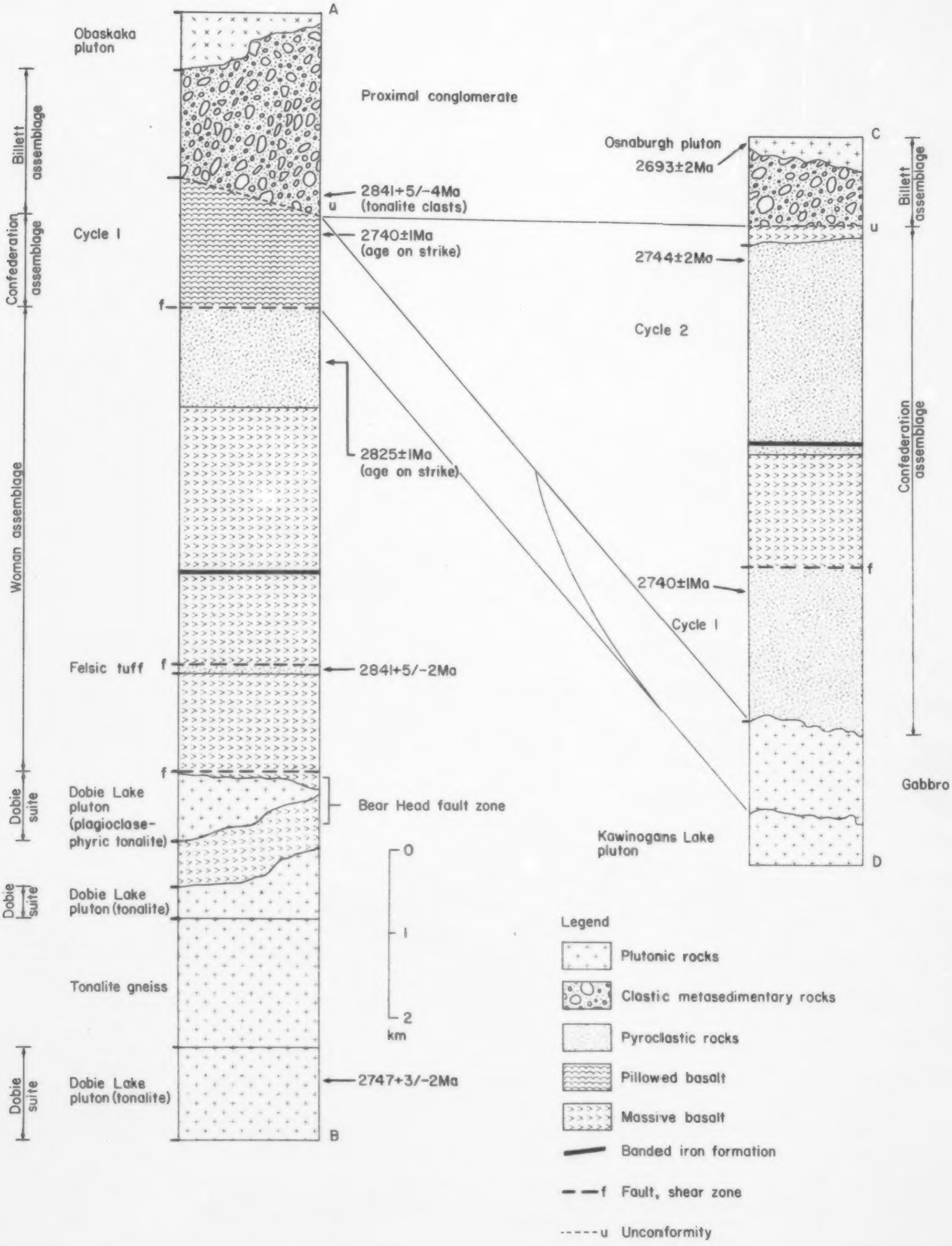

A
Obaskaka pluton
Billett assemblage
Proximal conglomerate
2841+5/-4Ma (tonalite clasts)
u
Confederation assemblage
Cycle I
2740±1Ma (age on strike)
f
2825±1Ma (age on strike)
Woman assemblage
Felsic tuff
f
2841+5/-2Ma
f
Dobie suite
Dobie Lake pluton (plagioclase-phyric tonalite)
Bear Head fault zone
Dobie suite
Dobie Lake pluton (tonalite)
Tonalite gneiss
Dobie suite
Dobie Lake pluton (tonalite)
2747+3/-2Ma
B
0
1
2
km
C
Osnaburgh pluton
2693±2Ma
Billett assemblage
u
2744±2Ma
Cycle 2
Confederation assemblage
f
2740±1Ma
Cycle I
Gabbro
Kawinogans Lake pluton
D
Legend
Plutonic rocks
Clastic metasedimentary rocks
Pyroclastic rocks
Pillowed basalt
Massive basalt
Banded iron formation
f Fault, shear zone
u Unconformity

Figure 6.11. Tectono-stratigraphic sections showing typical rock units and ages of the major assemblages in the Meen–Dempster greenstone belt. Locations of these sections are shown on Figure 6.9. (Age determinations are by F. Corfu, Royal Ontario Museum.)

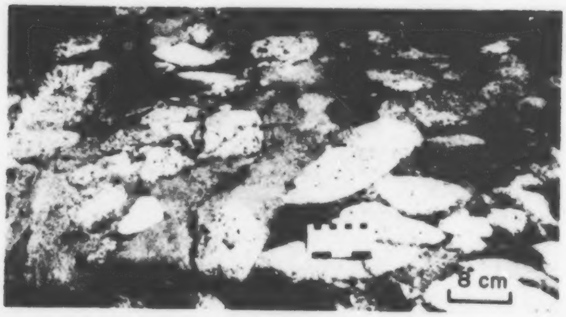


Figure 6.12. Cobble and boulder conglomerate, with interbeds of graded sandstone, on Billett Lake. The subangular clasts of trondhjemite petrographically resemble the Dobie Lake pluton and the U-Pb zircon age of 1 clast (2748^{+5}_{-4} Ma) corresponds to the age of the pluton. This conglomerate-sandstone sequence of the Billett assemblage lies adjacent to its northern contact with the underlying volcanic rocks of the Meen–Dempster greenstone belt.

nose of the anticline, the positioning of the anticline having been governed by the shape of the volcano.

The lower unit of the Confederation tholeiitic basalt flows is over 2000 m thick. The upper unit, 500 m thick, is composed of local rhyolitic flows near vents on the flanks of the main vent, but is dominated by rhyolitic to dacitic pyroclastic deposits of proximal vent facies, which grade to more distal, finer-grained deposits on the distal flanks of this volcano (Clifford 1969; Stott et al. 1987a, 1987b). The vent zone is marked by pervasive iron carbonate-filled vesicles in all rock types, local intense calcium carbonate alteration of basalt, silicification, abundant basaltic dikes and diatreme breccia pipes and local massive sulphide clasts in the coarser pyroclastic deposits.

The boundary of this assemblage with the St. Joseph assemblage (i.e., at the Cycle 1–Cycle 2 contact on Figure 6.6) is a disconformity; each assemblage sequence shares the same volcanic vent, the felsic flows and explosive products of which are well exposed among the islands on western Lake St. Joseph. The west boundary of Cycle 1 is a posttectonic granitic pluton that imposed contact strain and metamorphic aureoles upon the Confederation assemblage; substantial erosion of the Cycle 1 edifice apparently occurred before reactivation of the vent during St. Joseph volcanism (cycle 2); the west boundary of Cycle 3 with Cycle 2 is not clear and lies mainly within the lake, but is interpreted to be a fault on the basis of age determinations, with Cycle 3 slightly but significantly older than Cycle 2; the east boundary of Cycle 3 with Cycle 4 is not exposed.

Western Lake St. Joseph displays some of the best-preserved proximal and vent facies stratigraphy to be seen anywhere in the Superior Province (*see* Figures 6.2 and 6.3). The calc-alkalic edifice of the Confederation and Lake St. Joseph assemblages is built successively upon a tholeiitic basalt base and shows evidence in each assemblage of erosive products in the flanking basins as expected for a volcano that emerged above the water surface. This was interpreted by Clifford and McNutt (1971) as a composite stratovolcano; their study was the only detailed analysis of a volcano system in the Uchi Subprovince prior to Thurston's investigations in the central Birch–Uchi greenstone belt.

St. Joseph Assemblage (Volcanic)

The St. Joseph assemblage (*see* Figure 6.3) is composed of arc-like volcanic cycles of tholeiitic basalt and calc-alkalic andesite, dacite and rhyolite. Its most extensive exposure is in the vicinity of Lake St. Joseph and the southern half of the Miminiska–Fort Hope greenstone belt (*see* Figure 6.2). This is the youngest assemblage in the Uchi Subprovince and includes ages of calc-alkalic volcanism, in the upper formation of each cycle, ranging from about 2723 to 2713 Ma. This assemblage represents an episode of volcanism that postdates the 2750 to 2730 million-year-old Confederation assemblage in Ontario and its continuation in Manitoba as the 2730 million-year-old Bidou Lake Subgroup of the Rice Lake greenstone belt (Turek et al. 1989). The St. Joseph assemblage is the last episode of volcanism prior to the regional deformation that so dominates the southern half of the subprovince. In the eastern half of the Uchi Subprovince, the St. Joseph assemblage occupies the southern flank of the subprovince; it strikes locally (east of Lake St. Joseph) at a 30° angle to the southern boundary of the subprovince where the assemblage is terminated by the Sydney Lake–Lake St. Joseph Fault. The boundary of this assemblage with the adjacent Confederation assemblage marks a disconformable hiatus in volcanic activity in western Lake St. Joseph.

The basaltic base of this assemblage is overlain by a thick mantle of dacitic to andesitic pyroclastic deposits. This pyroclastic sequence is typified in the vicinity of Lake St. Joseph by an upward progression from dacitic, lithic tuff breccia, to highly vesicular tuff breccia, to an increasingly mafic dacitic to andesitic lapilli tuff. The coarsest pyroclastic deposits are to be found in the eastern part of Lake St. Joseph (Figure 6.13). This compositional change through the pyroclastic suite is consistent with eruption from a compositionally zoned magma chamber. The volume of preserved pyroclastic effusion in this final episode of Uchi volcanism (dated at about 2713 Ma; Corfu and Stott 1989) is volumetrically greater than similar deposits in nearby sequences of the Confederation assemblage.

Eagle Island Assemblage (Sedimentary)

In the vicinity of Eagle Island in western Lake St. Joseph, Clifford (1969) identified an assemblage of clastic and chemical sedimentary rocks that he called the Upper Clastic Rocks, which unconformably overlie the volcanic rocks. This sequence was subsequently mapped by Berger (1981), and described as a product of a submarine fan environment by Meyn and Palonen (1980). Basal deposits of eroded dacitic pyroclastic material lie unconformably upon the St. Joseph assemblage, the source of the detritus (Meyn and Palonen 1980; Berger 1981), and are succeeded upward by arenite- and wacke-sandstone beds, interbeds of mudstone, conglomerate and banded magnetite iron formation. Pebble clasts in the conglomerate have been successfully scavenged from arenite and iron formation beds, presumably deposited earlier in this submarine basin. An impressive range

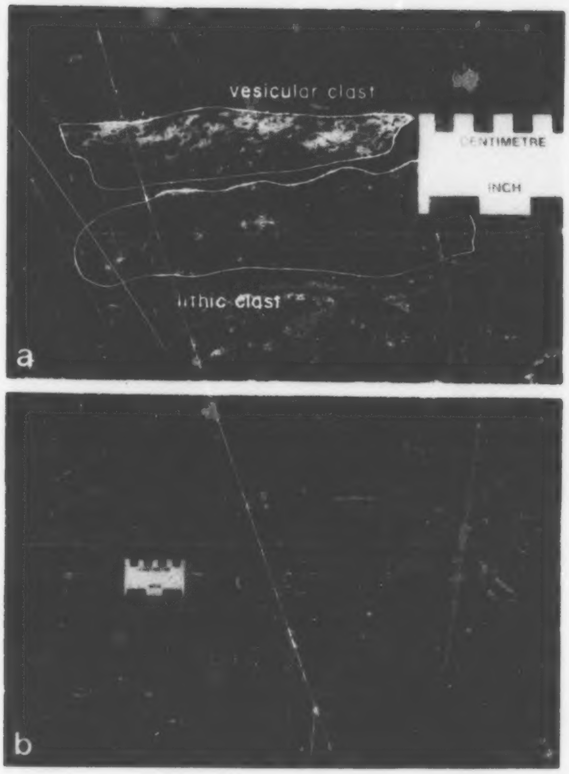


Figure 6.13. a) Dacitic pyroclastic breccia on eastern Lake St. Joseph in the uppermost calc-alkalic sequence of the St. Joseph assemblage. An age of 2713 Ma was obtained on a tuff stratigraphically above this outcrop. The pyroclastic breccia contains lithic and highly vesicular clasts in a more mafic matrix, with an increase in the presence of vesicular clasts upward in the unit. This unit is overlain by dacitic and andesitic tuff beds shown in Figure 6.13**b**. The tuffaceous beds become more dominantly mafic near the top of the sequence.

of sedimentary structures are preserved in this Eagle Island sequence. Meyn and Palonen (1980) have summarized the stratigraphy in a composite section shown in Figure 6.14. These authors favour an interpretation of this sedimentary assemblage, using various sedimentological criteria, as the product of a submarine fan environment.

There is sufficient evidence of slumping and dewatering to permit one to speculate that the earliest folds in the sedimentary rocks (Clifford 1969), which possess no associated cleavage, may be products of large-scale subsidence of the original depositional sequence. The subsidence may have been encouraged by the substantial thickness and density of iron formation beds in this unstable, clastic host and resulted in an abundance of dewatering features, sandstone diking, slump balls and slump folds. Subsequent regional folding would have tightened these folds and made their primary origin less obvious, creating a complexity of lineation orientations and refolded folds; this is well documented by Clifford (1969). Further discussion on the regional structures and the significance of the sedimentary Eagle Island assemblage is given in the structural geology section entitled Lake St. Joseph Greenstone Belt.

LANG LAKE GREENSTONE BELT

The Lang Lake greenstone belt (*see* Figure 6.2; Fenwick 1970, 1971; Fenwick and Srivastava 1972) is a typical composite of tholeiitic basalt flows and calc-alkalic pyroclastic deposits (*see* Table 6.6). A wacke-dominated sedimentary sequence at the east end of the belt contains banded magnetite iron formation beds; its relation to the rest of the belt has not been clearly defined. A magmatic age of 2749^{+8}_{-3} Ma (F. Corfu, unpublished data) has been determined for a dacitic tuff unit, in the lowermost part of this synclinally folded belt. Since the dacitic tuff lies in the stratigraphically lower section of the Lang Lake greenstone belt, the authors have provisionally included the strata, comprising this belt, with the Confederation assemblage. This is the oldest age for rocks assigned to the Confederation assemblage and is comparable to the age of a unit in the Red Lake greenstone belt.

MIMINISKA–FORT HOPE GREENSTONE BELT

The Miminiska–Fort Hope greenstone belt (*see* Figure 6.2) has been treated to only limited detailed government mapping by Wallace (1978, 1981a, 1981b), in the southern part of the belt, and by Prest (1944) in the northern part of the belt (referred to as the Keezhik Lake area here and located on Figure 6.2); reconnaissance mapping by Thurston and Carter (1970); and some detailed mapping by exploration companies. There are 5 tectono-stratigraphic subdivisions treated in this belt, as shown on Figure 6.3. Only the St. Joseph assemblage has been dated; the other subdivisions are based on structural and regional stratigraphic considerations, including potential correlations with other greenstone belts to the north and west. Each of these subdivisions is described below and features of 2 of the assemblages are summarized in Table 6.7.

Volcanic Rocks of Keezhik Lake Area

In the northern part of the belt, the authors have provided a provisional subdivision of strata into 3 unnamed assemblages (*see* Figure 6.3) based on stratigraphic and geophysical features. The northernmost assemblage, labelled 3 on Figure 6.3, is composed of basaltic flows with a marker unit of banded magnetite iron formation. This assemblage may correlate with the McGruer assemblage in the North Caribou greenstone belt (*see* Thurston, Osmani et al., this volume). It trends at a high angle to a second unnamed assemblage in the vicinity of Keezhik Lake (*see* Figure 6.2), labelled 4 on Figure 6.3. The possible correlation with the McGruer assemblage is based on similar rock types and particularly a magnetite iron formation unit forming a discontinuous aeromagnetic anomaly extending between the North Caribou greenstone belt and the northern Keezhik Lake area.

A comparison with the arrangement of assemblages in the Pickle Lake area (*see* Figure 6.3) leads one to suggest that the second "assemblage" may in future be shown to be a composite of 2 assemblages, correlative with the Pickle Crow and Woman assemblages of the Pickle Lake greenstone belt. This sequence of rocks faces consistently southward and is composed of massive to pillowed basalt flows with a marker unit of banded magnetite iron formation that may correspond to a prominent iron formation unit of the Pickle Crow assemblage on First Loon Lake in the Pickle Lake greenstone belt (*see* Figure 6.10b; Stott et al. 1989a, 1989b). This is succeeded southward by a basaltic pile supporting a unit of dacitic pyroclastic rocks accompanied by a quartz porphyry intrusion. The pyroclastic unit is speculated to be an extension of the Woman assemblage from the Pickle Lake greenstone belt.

A third unnamed assemblage in the northern part of the belt is labelled 4-5 on Figure 6.3 and is composed of basaltic flows and minor felsic volcanic units. It appears, west of Keezhik Lake, to lie unconformably upon a narrow sedimentary unit west of Keezhik Lake that lies at the top of the assemblage labelled 4 on Figure 6.3. This third assemblage also faces stratigraphically southward. Its southern contact with the St. Joseph assemblage, the fourth assemblage of this belt, is not well established, but is provisionally located where there is a regional change in the stratigraphic younging directions, from south facing amongst these

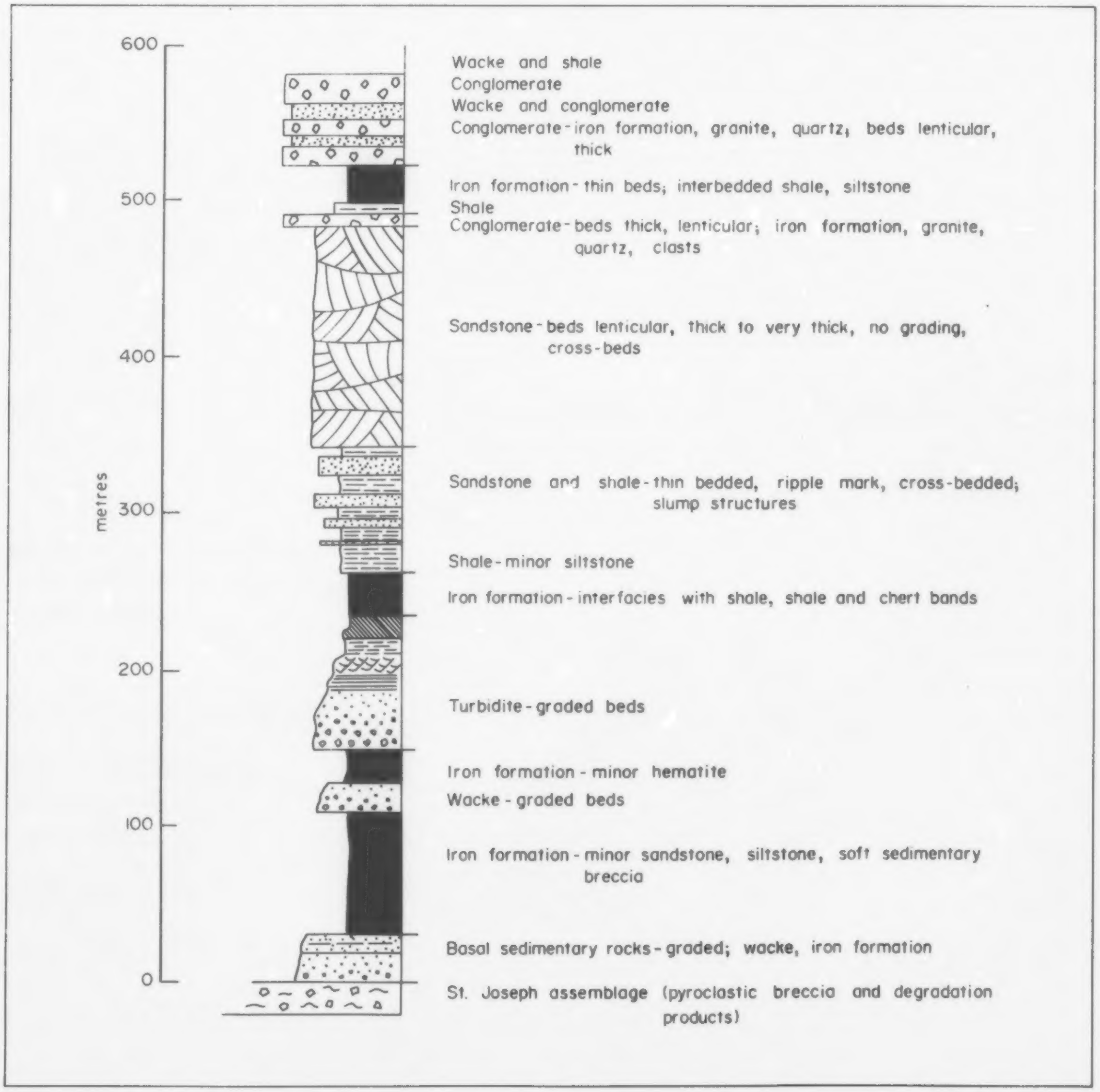

600
500
400
300
200
100
0
metres
Wacke and shale
Conglomerate
Wacke and conglomerate
Conglomerate - iron formation, granite, quartz, beds lenticular, thick
Iron formation - thin beds; interbedded shale, siltstone
Shale
Conglomerate - beds thick, lenticular; iron formation, granite, quartz, clasts
Sandstone - beds lenticular, thick to very thick, no grading, cross-beds
Sandstone and shale - thin bedded, ripple mark, cross-bedded; slump structures
Shale - minor siltstone
Iron formation - interfacies with shale, shale and chert bands
Turbidite - graded beds
Iron formation - minor hematite
Wacke - graded beds
Iron formation - minor sandstone, siltstone, soft sedimentary breccia
Basal sedimentary rocks - graded; wacke, iron formation
St. Joseph assemblage (pyroclastic breccia and degradation products)

Figure 6.14. Composite summary of the stratigraphy of the Eagle Island assemblage (*modified from* Meyn and Palonen 1980). The section is south facing.

northern rocks, to north facing amongst rocks that are clearly part of the St. Joseph assemblage in the vicinity of Miminiska Lake.

These interpreted subdivisions in the Keezhik Lake region are projected eastward into the northeastern region of the belt from aeromagnetic patterns. The northeasternmost region is dominated by basalt interbedded with some iron formation units (Thurston and Carter 1970) and is comparable with the lithological characteristics of the older units preserved locally in the northern half of Uchi Subprovince and in that part of the North Caribou terrane described farther to the north by Thurston, Osmani et al. (this volume).

St. Joseph Assemblage (Volcanic)

The southern half of the Miminiska–Fort Hope greenstone belt is interpreted to be dominated by young volcanic sequences of tholeiitic and calc-alkalic basaltic flows (Wallace 1981a) and overlain by calc-alkalic pyroclastic deposits ranging from andesitic to rhyolitic compositions. This range and the abundance of andesitic volcanic material corresponds with the uppermost cycle of the St. Joseph assemblage on Lake St. Joseph. Quartz-phyric felsic pyroclastic rocks at 3 locations in this assemblage (*see* Figure 6.3) show a range in age from 2723 to 2716 Ma (F. Corfu, unpublished data). This assemblage is interpreted to continue eastward, beyond Fort Hope, where it is interleaved on a regional scale with clastic sedimentary sequences in the eastern part of the belt (*see* Figure 6.3). The boundary between the St. Joseph assemblage and older volcanic rocks to the north is interpreted to project eastward beyond Fort Hope (*see* Figure 6.3), based on aeromagnetic interpretations only.

Sedimentary Rocks of Miminiska Lake Area

The sedimentary rocks of Miminiska Lake are composed of medial to distal turbidite wacke sediments and interbeds of banded magnetite iron formation. There is some folding of the beds, but with an overall northward sense of younging. The top of the assemblage is marked locally by conglomerate with a mix of volcanic, iron formation and granitoid clasts. Since there is uncertainty about the source of the detritus in this sequence and the original depositional setting of the assemblage, the sedimentary rocks are treated as an unnamed assemblage for the present. It is speculated that the assemblage evolved in its present position; it is just as conceivable that this sequence, which does not appear to be an original synclinal basin in the belt, was separated as a tectonic wedge from the northernmost part of the English River assemblage and transported northward to its present position. In this respect, the relationship of this sedimentary assemblage to the apparently interleaved volcanic and sedimentary sequences that characterize the east half of this belt (*see* Figure 6.3) merits an assessment of similarities to the northern part of the Quetico Subprovince east of Lake Nipigon, where Williams (1987, 1990) and Devaney and Williams (1989) documented features, in a similarly interlayered set of rocks, which are consistent with a fore-arc accretionary prism. The interlayering of strata in the easternmost Uchi Subprovince may correspond to a comparable pattern of interlayered volcanic and sedimentary units that characterizes the La Grande River Subprovince (Card and Ciesielski 1986) in Quebec (*see* Figure 6.1).

INTRUSIONS

Introduction

Most batholiths of the Uchi Subprovince are complexes of multiple intrusions of various ages, shapes and compositions. They represent different stages in the history of this subprovince. The granitoid batholiths outside of the greenstone belts have received only widely scattered mapping, but workers in recent years (e.g., Sage and Breaks 1982; Stott 1985, 1986; Noble 1989; Noble et al. 1985) have subdivided several of the irregularly shaped and elliptical batholiths into compositional suites based in part on aeromagnetic interpretation (*see* Map 2542, map case).

This section focusses on the characteristics of the batholiths and plutons—their 1) compositional variations and 2) timing of emplacement relative to the volcanic and regional tectonic events.

The fivefold compositional subdivision of intrusive rocks from the bedrock map of Ontario (see Map 2542, map case) classifies granitoid rocks which occur within the greenstone belts (i.e., the internal plutons) and outside the belts (i.e., the plutonic phases within the external batholiths). The legend subdivisions of the bedrock map distinguish the following compositional suites:

a) older gneissic and foliated tonalitic to granodioritic rocks that are typically synvolcanic or predate the Kenoran Orogeny, such as the Quarrier tonalite gneiss in the Seach–Achapi batholith (*see* Figure 6.10a)

b) younger syn- to posttectonic muscovite-bearing granitic rocks such as the Twiname stock (*see* Figure 6.6)

c) largely posttectonic massive granodiorite to granite, such as plutons of the Williams suite (*see* Figure 6.3)

d) rocks of the diorite-monzonite-granodiorite suite that are typically syn- to late Kenoran, and commonly occur near the Uchi–English River subprovince boundary, like the Doran Lake stock (*see* Figure 6.6)

e) mafic to ultramafic plutons that probably range in relative age from synvolcanic to posttectonic and are widely distributed within and near the margins of greenstone belts (*see* Figure 6.2)

Younger late to posttectonic granitic plutons have been identified as distinct intrusions within some of the external batholiths (*see* Figures 6.2 and 6.3); other batholiths, especially in the eastern third of the subprovince, east of Pickle Lake, remain unsubdivided and are provisionally included in the foliated tonalite suite.

In terms of relative timing of emplacement, the plutons are characterized as: "synvolcanic", "pre- to syntectonic" and "late to posttectonic", although such subdivisions have limited merit, as will be discussed later, since timing relative to tectonic events can only be tested on a local scale.

We shall review these characteristics by reference to 1) the typical features of several granitoid batholiths, their tectonic roles, and the variation of intrusive style (*see* Typical Features of Granitoid Batholiths); then 2) highlight some of the external and internal plutons in terms of their timing of emplacement relative to local magmatic and deformational events, and illustrate their role in the production of local strain fabrics upon the greenstone belts (*see* Relative Timing of Intrusions).

Typical Features of Granitoid Batholiths

Many batholiths are composed of a central mass of deformed and metamorphosed tonalite to granodiorite, rimmed by late tectonic, crescentic granodiorite plutons. The Trout Lake (Noble et al. 1985, 1989) and Lake St. Joseph batholiths (Stott 1985) are examples of this pattern. Some batholiths are dominated by late granodiorite plutons, such as the Fawcett Lake batholith southwest of the Meen–Dempster greenstone belt (Wallace 1983); some are dominated by single synvolcanic plutons, such as the Dobie Lake pluton north of the same belt (Stott and Wallace 1984). Still other batholiths contain large irregular shaped, stoping plutons with relatively limited depth extent (Szewczyk and West 1976) intruded into an unsubdivided mass of foliated to gneissic tonalite and granodiorite. Several examples of granitoid batholiths are described below to illustrate the most typical features.

LAKE ST. JOSEPH BATHOLITH

The Lake St. Joseph batholith (*see* Figure 6.2) is an example of the structural and lithological complexity of batholiths and an example of the long history of thermal and structural reactivation of its plutonic components. Parts of the Lake St. Joseph batholith (*see* Figure 6.6) were initially mapped by Goodwin (1965) and Clifford (1969). The batholith was further subdivided into its component parts by Stott et al. (1987a, 1987b). This batholith illustrates some basic features observed in several of the larger batholiths common to granite-greenstone terranes (cf., Schwerdtner et al. (1979) who described these features in batholiths of the Wabigoon Subprovince).

1. The Lake St. Joseph batholith is subdivided into areas of older, metamorphosed and tectonized tonalite to granodiorite (the multi-phase Pembina Tonalite Gneiss and the homogeneous foliated Searson Tonalite) and younger plutons of granodiorite to monzogranite that contain strain fabrics formed during plutonic emplacement. The younger plutons (Osnaburgh pluton, Carling pluton and Twiname stock) display little or no evidence of metamorphism.
2. The older tonalite phases listed above comprise the central core of the batholith. The oldest tonalite phase in the Pembina Tonalite Gneiss (ca. 2887 Ma) occurs with other tonalitic phases, cut by several sets of tonalite and granite dikes. The structural complexity and the record of multiple intrusive phases contrast markedly with the younger plutons, which contain fewer rock types and very weak planar fabrics attributed to their original emplacement.
3. Long slivers of volcanic and clastic sedimentary rocks form narrow screens that subdivide parts of the tonalitic core. Other supracrustal slivers form discontinuous screens between the tonalite and the younger plutons. Such features are similar to those described by Schwerdtner et al. (1985) in the Wabigoon Subprovince.
4. The younger granodiorite-monzogranite plutons are tabular to crescent shaped and separate the tonalitic core from the surrounding volcanic belts. They appear to have been injected along the interface between an older tonalitic terrane and the surrounding greenstone belt, and resemble sheet intrusions, comparable to those described by Schwerdtner et al. (1983) in the central part of the Wabigoon Subprovince. These biotite-bearing plutons (e.g., the Osnaburgh and Carling plutons) are petrographically similar and possess varying proportions of potassium feldspar megacrysts. In places, the tabular megacrysts are sufficiently abundant to form clear primary flow textures. These plutons are generally free of supracrustal inclusions.
5. The young, posttectonic, peraluminous Twiname stock (*see* Figure 6.2) is a comparatively large example of a suite of typically small pegmatite stocks and dikes that occur in the southern half of the Uchi Subprovince. This stock probably bears a temporal and genetic relation to the peraluminous granite plutons in and adjacent to the English River Subprovince. The Twiname stock is included in the Churchill suite of muscovite-bearing granite plutons and is described in the English River Subprovince chapter by Breaks (this volume).
6. The late to posttectonic plutons imposed contact strain aureoles (defined *in* Tectonic Elements—Structures and Metamorphism) on both the surrounding volcanic belt and the tonalitic core. The imposed strain is recognized most particularly in adjacent volcanic rocks by steeply plunging lineations of prismatic amphibole crystals and mineral aggregates within the aureole; these lineations are parallel to the magmatic stretching lineations in the pluton, and are accompanied by a contact thermal aureole imposed on the supracrustal envelope.

Another batholith to the east of the Lake St. Joseph batholith shows similar features as described below.

SEACH–ACHAPI BATHOLITH

The Seach–Achapi batholith (*see* Figure 6.2) contains early sodic and younger potassic plutons, like the Lake St. Joseph batholith, but it also shows evidence that plutonism in the northern Uchi Subprovince is synchronous locally with volcanism in the southern Uchi Subprovince. The western part of the Seach–Achapi batholith consists of several lithologic components (Stott et al. 1989b) as shown in Figures 6.2 and 6.10. The central tonalitic core (Quarrier tonalite gneiss) has yielded an age at 1 locality of 2821 Ma; it is metamorphosed and recrystallized and more strongly deformed near its contacts with younger plutons, which include the Second Loon pluton of biotite-hornblende granodiorite and the Kagami pluton of quartz-phyric granodiorite (*see* Figure 6.10). Both plutons are late intrusions and the Second Loon pluton appears to have imposed a contact strain aureole upon the adjacent Pickle Lake greenstone belt (to be described later; *see* Figure 6.10). But this pluton, viewed as late to posttectonic in the northern part of the Uchi Subprovince, is dated at 2713±2 Ma (F. Corfu, unpublished data), and thus also corresponds in age to the youngest volcanism of the St. Joseph assemblage forming the southern edge of the Uchi Subprovince (*see* Figure 6.6). Indeed, there are several examples of syn- to posttectonic plutons, intruded into the northern half of the central Uchi Subprovince, which are synchronous with volcanism farther south (Corfu and Stott 1989); these include the 2736 million-year-old Hammerton Lake pluton and 2722 million-year-old Kawashe Lake stock (*see* Figure 6.9); and the 2741 million-year-old Ochig Lake pluton and Pickle Lake stock (*see* Figure 6.10; F. Corfu, unpublished data). As shown earlier by phases of the Lake St. Joseph batholith, there is an episodicity of magmatism in this region, although the nature of the link between the volcanism and plutonism remains speculative, and will be discussed later.

TROUT LAKE BATHOLITH

The Trout Lake batholith (*see* Figure 6.2), like the previous 2 batholiths, contains an older core of foliated tonalite, containing remnants of tonalitic gneiss, surrounded by younger granitic to granodioritic plutons.

Noble (1989) and Noble et al. (1985) have shown that the core of this batholith is composed of various units of foliated to gneissic tonalite, partly surrounded to the south and west by a crescent-like mass of younger granite to granodiorite adjacent to the greenstone belts (*see* Figures 6.2 and 6.3). Noble has determined U-Pb zircon ages of 2 phases of foliated tonalite to be 2838^{+4}_{-3} and 2806^{+12}_{-2} Ma, respectively, and the crescentic Walsh Lake pluton, east of the Red Lake greenstone belt, to be 2699±1 Ma in age (Noble 1989; Noble et al. 1989). This younger pluton has produced a broad contact metamorphic aureole in the eastern part of the Red Lake greenstone belt (Andrews et al. 1986). The most critical features of this batholith are the tonalite phases. The oldest foliated tonalite phase intrudes the base of the Balmer assemblage where it is transected by the batholith. This 2838 million-year-old tonalite phase is synchronous with *circa* 2840 million-year-old felsic volcanic rocks of the Woman assemblage overlying the Balmer assemblage. The strata-parallel boundary between the Balmer and Woman assemblages is an age disconformity. Although a genetic relationship has not been demonstrated geochemically between the tonalite intrusion and the Woman assemblage, this spatial and timing relationship implies that the Woman assemblage was either developed in place upon the older Balmer assemblage or was tectonically transported only a short distance. In contrast to this interpretation is the position taken by Thurston, Osmani et al. (this volume) who advocate the 2806 million-year-old main tonalite phase, which Noble has shown to straddle the boundary between the Balmer and Woman assemblages, may be a stitching pluton; this would provide a minimum age of accretion of these 2 assemblages, which in this scenario, are parts of 2 separate terranes.

BERENS RIVER SUBPROVINCE

The Berens River Subprovince (as defined by Card and Ciesielski 1986) is a tabular region dominated by plutonic rocks. The geology of this region is described by Thurston, Osmani et al. (this volume). Here we examine briefly the relationship between the Berens River Subprovince and the Uchi Subprovince. The southern flank of the Berens River Subprovince merges into the northern Uchi Subprovince without any evidence of a boundary. The lack of a boundary is concluded from the following:

1. there is no fault between the Uchi Subprovince and the broad plutonic belt to the north that forms the Berens River Subprovince (*see* Map 2542, map case)
2. there is a northward continuity of strike of the volcanic strata, confirmed by tracing aeromagnetic anomalies and discontinuous slivers of supracrustal rock, from the Red Lake greenstone belt into the Berens River Subprovince, towards the McInnes Lake and Hornby Lake greenstone belts, which lie 100 km north of the Red Lake greenstone belt (Stone 1990)
3. in the McInnes Lake and Hornby Lake greenstone belts, there are 2.9 billion-year-old ages of volcanism (D. Davis, Royal Ontario Museum, personal communication, 1991) that are synchronous with the volcanism in parts of the western Uchi Subprovince
4. there is a continuity of plutonic intrusions across the Uchi–Berens River subprovince boundary (Stone 1988, 1990)

This evidence supports the interpretation advanced in Thurston, Osmani et al. (this volume) that the Berens River Subprovince represents a deeper midcrustal exposure of the granite-greenstone terrane, which includes the older assemblages in the north Uchi Subprovince and much of the southern Sachigo Subprovince. This terrane, called the North Caribou greenstone terrane, is described further in a terrane analysis of the northwestern Superior Province by Thurston, Osmani et al. (this volume) and later in this chapter.

The broad plutonic belt that forms the Berens River Subprovince contains the compositional suites of foliated tonalite to granodiorite, tonalite gneiss, with minor supracrustal inclusions, and a volumetrically predominant suite of late, *circa* 2705 to 2690 million-year-old plutons (Corfu and Ayers 1984; F. Corfu, unpublished data). Most of these late plutons are exceptionally large granite to granodiorite bodies, typically with megacrystic potassium-feldspar and with such an abundance of magnetite grains as to highlight these intrusions with an anomalous magnetic susceptibility visible on aeromagnetic maps. The plutons imposed little contact strain upon the surrounding rocks and stoped their way into place.

This is such a pronounced plutonic belt with components that probably reflect a long history of plutonism, yet with a history that culminates in a large volume of magmatism at the time of the Kenoran Orogeny, *circa* 2700 Ma. It parallels the Uchi–English River subprovince boundary; it straddles and exposes a deeper crustal section of the North Caribou terrane; and its size and tabular shape, which are similar to Proterozoic and Mesozoic magmatic arcs (cf., Hoffman 1990; Pitcher et al. 1985) lead one to speculate that the late plutonism and crustal exhumation of this belt had developed during a major thermal event, arising from northward subduction of Archean lithosphere and collision along the southern flank of the Uchi–Sachigo superterrane.

Relative Timing of Intrusions

Identifying the timing of pluton emplacement relative to other magmatic and deformational events is controversial and dependent upon unequivocal criteria that are not always forthcoming (cf., Paterson et al. 1989). The attempt to so characterize some of the intrusions in the Uchi Subprovince is dependent upon: 1) the availability of high-precision U-Pb age determinations; 2) distinguishing unmetamorphosed plutons with magmatic foliations and lineations that contrast with the fabric of metamorphosed rock units; 3) distinguishing contact metamorphic aureoles imposed by some plutons, which postdate larger scale domains of strain observed in the greenstone belts; 4) identifying a regional strain pattern which is expressed on a scale exceeding the size of an individual pluton that carries the same fabric orientation; and 5) other regional-scale field observations that impose limits on timing relationships amongst rock units.

SYNVOLCANIC INTRUSIONS

Within the greenstone belts, synvolcanic and late to posttectonic felsic intrusions can locally be distinguished, but such labels should be applied with caution, because they are appropriate only to the local time-space framework and in reference to a specific tectonic event or an orogenic period. An example of the paradoxical nature of these labels is given later in this section. Mafic to ultramafic intrusions are less easily dated by the U-Pb technique and it is not always clearly determined that the intrusions are synvolcanic. Synvolcanic timing is more evident among thin sills within the volcanic pile (e.g., Raudsepp and Ayres 1982), but their relative timing can be surprisingly late in the tectonic record. These mafic intrusions are therefore discussed as pre- to syntectonic intrusions in the section entitled Pre- to Syntectonic Plutons.

The felsic synvolcanic intrusions range in composition from trondhjemite to quartz diorite to granodiorite. The intrusions include subvolcanic, quartz and feldspar porphyry stocks, dikes and sills; the most prominent example is a suite of elongate bodies parallel to the regional stratification within the Confederation assemblage of the Birch–Uchi greenstone belt (Thurston 1985b). These quartz and plagioclase-phyric intrusions are characterized by substantial granophyric intergrowths. The rocks are deformed and metamorphosed and display aggregates of chlorite and sericite. Others include equigranular to quartz porphyritic stocks having an irregular to subcircular shape. Some of these bodies have been dated and the ages correspond to those determined for the enveloping volcanic assemblage.

An example, in the Birch–Uchi greenstone belt, is an irregularly shaped, metamorphosed, subvolcanic quartz diorite intrusion—the Found Lake stock (2741^{+1}_{-2} Ma, F. Corfu, unpublished data; *see* Figure 6.5)—which is synchronous with extrusive rocks of the Confederation assemblage. A genetic link between these intrusions and the extrusive rocks is confirmed by the closely comparable whole rock and trace element geochemistry of granodiorite porphyries and felsic volcanic rocks of the Confederation assemblage (Thurston and Fryer 1983). Porphyry intrusions are commonly associated with calc-alkalic volcanic centres; smaller intrusions of this type are present near volcanic centres on Meen Lake in the Meen–Dempster greenstone belt (Stott and LaRocque 1983a), in the Pickle Crow assemblage near Pickle Lake (Ferguson 1966) and in the Keezhik Lake area of the northern Miminiska–Fort Hope greenstone belt (Prest 1939).

The next example of a deformed and metamorphosed synvolcanic intrusion is the North Bamaji pluton which is over 30 km long, adjacent to the Bamaji–Blackstone batholith (*see* Figure 6.2); it is a high-level, subvolcanic body of very fine-grained, slightly porphyritic trondhjemite and is strongly deformed by regional flattening and localized shearing. The stock crystallized synchronously (ca. 2806 Ma; Scharer 1989) with the extrusion of the adjacent volcanic rocks of the Wonnan assemblage. It is host to concentrations of molybdenite, which are present only in a few mineralized copper-molybdenum porphyry intrusions and have been documented in the Superior Province and reviewed by Ayres and Cerny (1982). A similar example of a mineralized porphyry system has been documented farther north in the Lang Lake greenstone belt by Findlay and Ayres (1977).

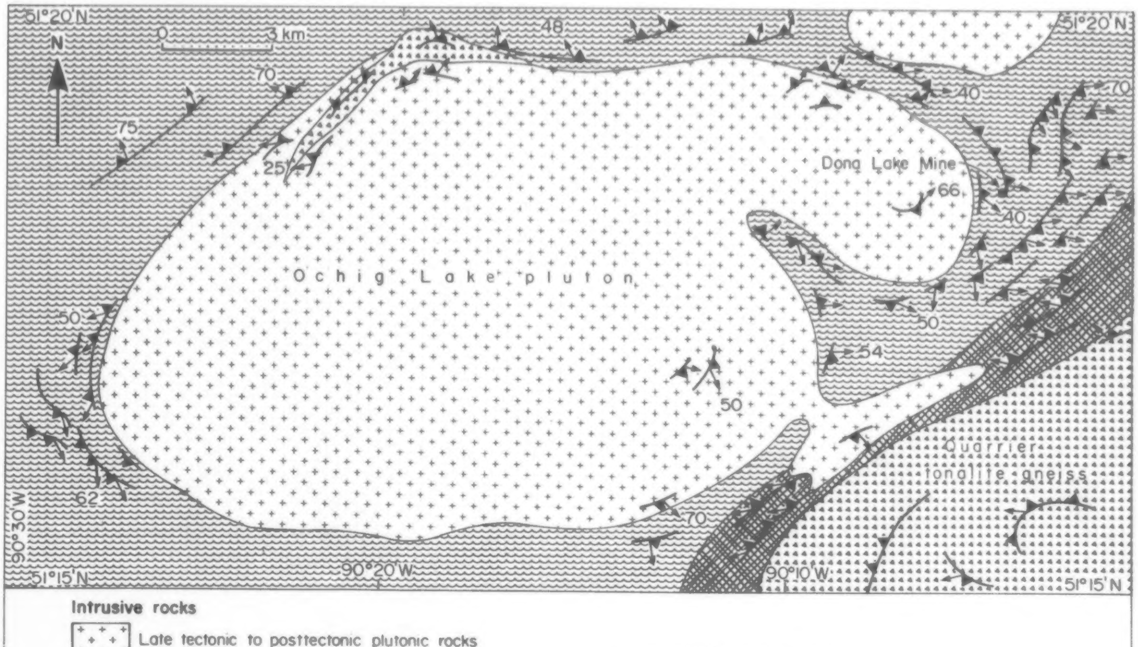
The Dobie Lake pluton (*see* Figure 6.2), north of the Meen–Dempster greenstone belt, is an example of a less clearly metamorphosed and orogenically deformed synvolcanic intrusion. It is a hornblende-biotite tonalite body, 40 km long, with a shallowly dipping domical core and a low magnetic susceptibility that reveals its presence on an aeromagnetic map, surrounded by greenstone belts and tonalite gneiss. Its age (2747 Ma) is comparable with that of a felsic volcanic unit (2749 Ma) in the southern part of the adjacent Lang Lake greenstone belt. Although it has imposed a narrow contact metamorphic aureole upon the adjacent greenstone belts, structural and paleomagnetic evidence show that it has been treated to subsequent regional orogenic shortening and rotation (Stott and Wilson 1986b; Hale and Lloyd 1989). The pluton is weakly metamorphosed and notably it carries a strain fabric with a stretch lineation near its margins, which is attributed to regional orogenic shortening (*see* the Summary in the section entitled Tectonic Elements—Structures and Metamorphism and Figure 6.9).

Less obvious as "synvolcanic plutons" are the trondhjemite to granodiorite bodies in the Pickle Lake greenstone belt. The Ochig Lake pluton and Pickle Lake stock (*see* Figure 6.10a) crystallized at about 2740 Ma. The Hooker–Burkoski stock is similar to the sodic Ochig Lake pluton in composition, fabric and relationship to surrounding rocks. These bodies are similar in age to the felsic extrusive rocks of the Confederation assemblage in the Meen–Dempster greenstone belt nearby. The synvolcanic relationship, however, is not obvious for 2 reasons: 1) the plutons do not intrude volcanic rocks of the same age, but are emplaced farther to the north into older crust and are not deformed by any regional shortening; 2) they are subcircular intrusions that produced contact strain and thermal aureoles upon their supracrustal host. The Ochig Lake pluton (Figure 6.15) postdates the regional deformation in the Pickle Lake greenstone belt and is, on a local scale, a posttectonic intrusion, but it is synvolcanic with respect to the Confederation assemblage and pre-tectonic with respect to the main deformational expression of the Kenoran Orogeny recorded in the southern half of the belt. The tectonic significance of this pluton is further described in the structural geology section entitled Contact Strain Aureoles.

PRE- TO SYNTECTONIC PLUTONS

Pre- to syntectonic plutons include: 1) foliated to gneissic tonalite masses in the external batholiths; 2) discrete synvolcanic plutons described earlier (*see* Synvolcanic Intrusions); 3) some evidently syntectonic plutons that are not related to volcanic episodes nearby; and 4) undated and variably synvolcanic to syntectonic mafic and ultramafic intrusions. The following are examples of each type.

1. Tonalite gneiss in the external batholithic complexes is restricted to either thin slivers, preserved near greenstone belt margins such as along the northern Meen–Dempster greenstone belt boundary, or as central cores in the batholiths as in the Lake St. Joseph and Trout Lake batholiths (*see* Figure 6.2). These rocks have not been mapped in detail and few have been dated, but much of the gneiss appears to be concentrated in the northern half of the subprovince. Foliated tonalite bodies, like the Dobie Lake pluton, are prime candidates for classification as synvolcanic intrusions, but geochemical and geochronological correlations have yet to be tested.
2. Intrusions that have close spatial and temporal relationships with volcanism are: hypabyssal porphyries abundant in the Birch–Uchi greenstone belt (Thurston 1985b), for example, and in the Lang Lake greenstone belt (Findlay and Ayres 1977); and larger fine-grained bodies, like the North Bamaji pluton (Wallace 1985), described in the previous section.


51°20'N
0
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48
70
75
25
51°20'N
40
70
Dong Lake Mine
66
40
Ochig Lake pluton
50
50
54
50
62
70
Quarrier tonalite gneiss
90°30'W
51°15'N
90°20'W
90°10'W
51°15'N
Intrusive rocks
Late tectonic to posttectonic plutonic rocks
Pretectonic to syntectonic gneiss
Volcanic rocks
Contact strain domain imposed by late plutons
Early deformation domain
Producing gold mine
Schistosity
50 Mineral elongation and mineral aggregate shape lineation; plunge and direction
Geological boundary

Figure 6.15. The Ochig Lake pluton is a 2741 million-year-old trondhjemitic intrusion in the Pickle Lake greenstone belt (*see* Figure 6.2). Its internal schistosity and quartz-aggregate lineations are oriented consistent with a domical structure. The volcanic envelope around the pluton has been deformed by the emplacement of the pluton and shows orientations of schistosity and lineations that parallel the magmatic strain fabric in the pluton.

3. Some examples of syntectonic plutons not related to volcanism nearby can be seen in the vicinity of the Red Lake greenstone belt, where several episodes of felsic plutonism at about 2730, 2720 and 2700 Ma have been documented by Corfu and Andrews (1987). Within the belt, the McKenzie and Dome stocks (*see* Figure 6.4) were intruded at about 2720 and 2718 Ma, respectively. These rocks appear to postdate the Confederation assemblage rocks in the southern part of the Red Lake greenstone belt, although they are synchronous with St. Joseph assemblage volcanism in the eastern part of the subprovince. These stocks have been interpreted as late tectonic intrusions by Corfu and Wallace (1986). The Dome stock is composed of biotite and hornblende granodiorite to trondhjemite; it postdates local isoclinal folding within the belt and yet is affected by local auriferous shear zones. It intruded at the junction between several volcanic assemblages and discordantly cuts the boundary between Balmer and Confederation assemblages; it therefore limits the tectonic juxtaposition of these and other assemblages to a minimum age of 2718 my.
4. The time relation of mafic and ultramafic intrusions to volcanic and tectonic episodes is less well constrained at present because of the scarcity of minerals that can be dated by the U-Pb technique. The wider application in future of U-Pb age determinations with baddelyite should provide the needed age constraints in mafic rocks.

In general, mafic to ultramafic intrusions are elongate and are concordant or slightly discordant to the volcanic strata enveloping them. A few are differentiated and sill-like, but most bodies are composed of homogeneous gabbro and their resistance to penetrative regional deformation depends on their size and shape. All intrusions have been deformed along with their volcanic hosts but some, such as the Kawashe gabbro (*see* Figure 6.9), have more successfully restricted much of the strain to their outer margins or to discrete shear zones within themselves.

Gabbroic plutons, found in most greenstone belts, locally crosscut assemblage or terrane boundaries; some appear to be older and potentially more useful for tectonic analysis as stitching plutons than the more prolific late to post–Kenoran granodiorite stocks emplaced typically at 2700 Ma. For example, the Kawashe gabbro in the Meen–Dempster greenstone belt is a subcircular intrusion that straddles the boundary between the Woman and Confederation assemblages. It postdates the 2740 million-year-old Confederation volcanic rocks and is intruded by the 2722 million-year-old Kawashe stock. Future determination of its age would provide a more precise minimum age constraint to the formation of that boundary.

Mafic and ultramafic intrusions are assumed to be synvolcanic, but it is also recognized that late to posttectonic mafic stocks and separate mafic phases of larger granitic plutons have been described elsewhere (e.g., in the Wabigoon Subprovince, Sutcliffe et al. 1990). They occur not only within the greenstone belts, but also among the poorly described mafic bodies in the granitic batholiths of the northern Uchi Subprovince. Two examples of mafic and ultramafic intrusions are described, one a layered intrusion and the other an irregularly shaped gabbro body with local concentrations of anorthosite.

The western Lake St. Joseph sill (*see* Figure 6.6) is a slightly discordant intrusion up to 500 m thick and extends about 11 km along strike with the volcanic strata. It transects the uppermost exposure of the Confederation assemblage and, according to Smith (1977) and Berger (1981), its magmatic layering and top direction conform with the northward-facing assemblage stratigraphy. Differentiation is displayed by minor dunite metamorphosed to an olivine pseudomorph assemblage (serpentine-magnetite-calcite-chlorite), succeeded by more voluminous pyroxenite and gabbro with abundant pegmatitic clots. The stratigraphic top direction was determined by Smith (1977) from several features including upward-decreasing anorthite content of the plagioclase and increasing quartz content, cross-stratification, and greater concentrations of less dense felsic minerals in the upper parts of the pegmatite clots.

In the Meen–Dempster greenstone belt, there are several gabbroic masses that discordantly intrude the volcanic rocks. An example is the Dempster gabbro (*see* Figure 6.9), an irregularly shaped body that contains a narrow anorthosite phase on Dempster Lake. It is regionally deformed and metamorphosed with a long axis parallel to the stratigraphic layering. Deformation in the gabbro is variable with zones of intense shear taking up a substantial share of the bulk penetrative strain. It intrudes the 2740 million-year-old Confederation assemblage and contains inclusions of basalt and only partly assimilated banded magnetite layers that have not been rotated. More significantly, the Dempster gabbro also intrudes the sedimentary Billett assemblage and contains xenoliths of that assemblage. Since the Billett assemblage unconformably overlies the volcanic sequences of the Confederation assemblage, the Dempster gabbro is clearly a late intrusion, but predates the regional D_1 deformation described in the Summary of the section entitled Structures. Its value as a time marker is evident and exemplifies the tectonic-time constraining potential of other magmatic events.

LATE TO POSTTECTONIC PLUTONS

Reconnaissance mapping of the Uchi Subprovince has shown that the granitoid complexes of the subprovince are dominated by late to posttectonic plutons. These are largely represented on the map of bedrock geology in Ontario (*see* Map 2542, map case) as massive granodiorite to granite. It has been determined that most of these bodies were likely to have been emplaced during the interval 2690 to 2705 Ma, based on their crystallization ages, similar lithologic characteristics and contact relations with adjacent rock units (*see* Map 2576, map case). In the central part of the Uchi Subprovince, these late intrusions are assigned to the Williams suite (*see* Figure 6.3); they are typically biotite-hornblende granodiorite to monzogranite. Some of the intrusions, notably those members of the suite in the Berens River Subprovince, are characterized by an anomalous abundance of magnetite. The large monzogranite pluton centred on Williams Lake in the Berens River Subprovince, northwest of Pickle Lake, is at least 20 by 80 km in size. It possesses variable abundances of potassium-feldspar megacrysts and fine-grained magnetite. Its size, relative homogeneity, its stoped boundaries with the surrounding rocks and the lack of a strain envelope resemble the features of other voluminous, late plutons in the Wabigoon Subprovince (e.g., White Otter pluton and Indian Lake pluton) and Winnipeg River Subprovince (e.g., Lount Lake batholith); the Indian Lake pluton north of Ignace has been shown by gravity modelling to be only 2 km thick (Szewczyk and West 1975) and the Williams Lake batholith is probably pancake-like as well.

Numerous late plutons have been found in recent years to have intruded along greenstone-granite boundaries and formed curved sheet-like crescents or broader wing-shaped bodies in plan view. Such plutons have been identified within the external batholiths (some examples were mentioned in the Introduction to this section). The features of one example will be highlighted. The Kagami pluton in the Seach–Achapi batholith (*see* Figures 6.10 and 6.3) is similar to crescent-shaped intrusions elsewhere; it is slightly arcuate in plan, or "boot-shaped" and appears to have invaded a pre-existing boundary between the greenstone belt and the Quarrier tonalite gneiss, and deformed the adjacent rocks during its emplacement. The following is taken from a reconnaissance strain analysis and microtextural study by Green (1987). This pluton is dominated by a single, homogeneous phase of granodiorite with quartz aggregates 2 to 20 mm in length (Figures 6.16a and 6.16b), and a finer-grained, plagioclase-phyric border phase, which is similar in modal composition. Mapping of the planar and linear alignment of minerals, especially quartz aggregates, within the pluton indicates that the strain fabrics were produced by the emplacement of the pluton and not by a regional deformation event. The fabric pattern within the pluton plus the pluton shape and its location at a greenstone-gneiss contact are comparable to crescentic plutons observed elsewhere in the Superior Province by Schwerdtner et al. (1983). In particular, the Kagami pluton is similar to the Jackfish Lake–Weller Lake pluton in Wabigoon Subprovince, a crescentic pluton described by Schwerdtner et al. (1983) as characterized by: 1) discordant relationships with adjacent rock units; 2) a prolate fabric with subhorizontal extension lineations converging towards the hinge line or plane of symmetry of the crescent; 3) at least 1 parasitic lobe showing increased flattening strain, with the Kagami pluton displaying a more evolved parasitic dome centred on its northern arm and probably developed at a late stage of pluton emplacement; 4) schistosity attitudes at the margins of the pluton that are consistent with the shape of true synformal sheets, but locally consistent with antiformal sheets where the pluton dips inward like a funnel.

Figure 6.16a. Typical exposure of the Kagami pluton in the Seach–Achapi batholith. The pluton is composed of medium-grained granodiorite with quartz aggregates suitable for determining orientation and magnitude of the late strain increment during final pluton emplacement and cooling.

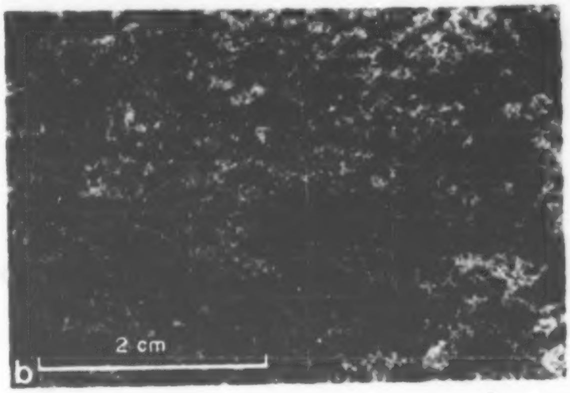


Figure 6.16b. Ellipsoidal quartz aggregate, approximately 2 cm in length, in a medium-grained homogeneous matrix. Planar fabric is weak in places compared to the lineation defined by the long axes of quartz aggregates.

Examples of late to posttectonic plutons within the greenstone belts are described as: typically subcircular in plan, with narrow penetrative contact strain aureoles, but imposing a broad "passive" deflection of the surrounding stratigraphy to make room. Most of these plutons are hornblende-biotite granite to granodiorite and some bear marginal syenite phases. A few plutons, such as the Doran Lake stock south of Lake St. Joseph (*see* Figure 6.6), show petrographic similarities to hornblende granodiorite to diorite bodies of the diorite-monzonite-granodiorite suite of the Shebandowan greenstone belt near Thunder Bay; this suite is postulated by Stern et al. (1989) to be associated with a mantle-derived sanukitoid suite. A number of these plutons appear to be aligned along the southern margin of the Uchi Subprovince (*see* Map 2542, map case).

Also late in the magmatic record are muscovite-bearing granite bodies which are typically small dikes and pods within the greenstone belt situated close to the Uchi–English River subprovince boundary. But some, such as the Twiname stock in the Lake St. Joseph batholith are sizeable. This stock is apparently related to small pegmatite pods within the greenstone belt that show anomalous rare-element concentrations as discussed in the section entitled Mineral Deposits. The Twiname stock is approximately 8 to 12 km in diameter and is composed of coarse tourmaline-bearing muscovite granite pegmatite with aplitic, garnetiferous, granite dikes. The surrounding granitic rocks

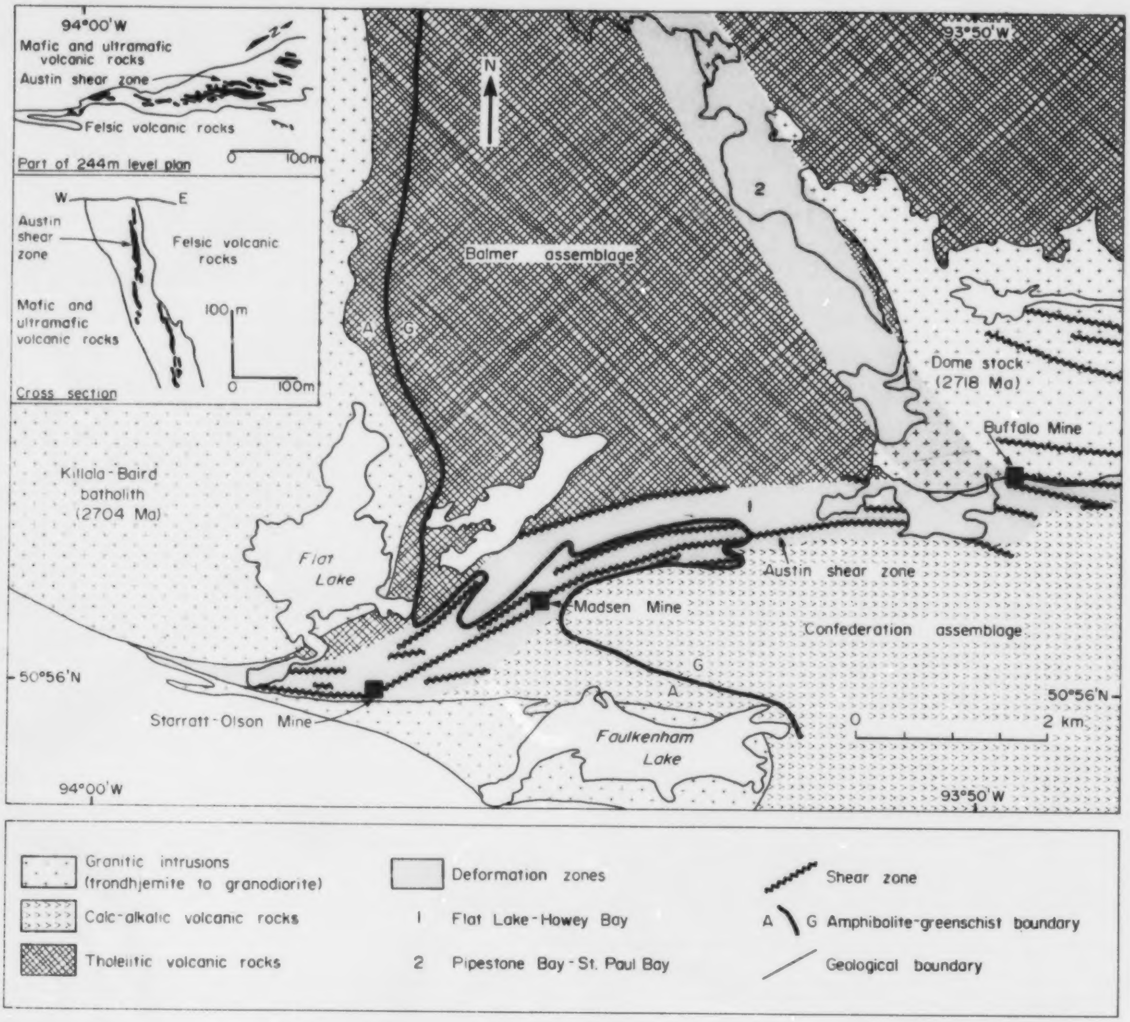

94°00'W
Mafic and ultramafic volcanic rocks
Austin shear zone
Felsic volcanic rocks
Part of 244m level plan
0
100m
W
E
Austin shear zone
Felsic volcanic rocks
Mafic and ultramafic volcanic rocks
100 m
0
100m
Cross section
N
93°50'W
2
Balmer assemblage
A
G
Dome stock (2718 Ma)
Buffalo Mine
Killala-Baird batholith (2704 Ma)
1
Flat Lake
Austin shear zone
Madsen Mine
Confederation assemblage
G
A
50°56'N
50°56'N
Starratt-Olson Mine
Faulkenham Lake
0
2 km
94°00'W
93°50'W
Granitic intrusions (trondhjemite to granodiorite)
Calc-alkalic volcanic rocks
Tholeiitic volcanic rocks
Deformation zones
1 Flat Lake-Howey Bay
2 Pipestone Bay-St. Paul Bay
Shear zone
A G Amphibolite-greenschist boundary
Geological boundary

Figure 6.17a. Map of part of the Red Lake greenstone belt, *modified from* Andrews et al. (1986), to show the deformation zone that follows the boundary between the Balmer and Confederation assemblages. The tectonic boundary is cut by the 2718 million-year-old Dome stock and reactivated to form the Flat Lake–Howey Bay deformation zone, which was overprinted by a thermal aureole from the 2704 million-year-old Killala–Baird batholith. The deformation zones contain a greater than normal abundance of discrete shear zones than is normal in the rest of the belt; the shear zones are interpreted by Andrews et al. (1986) to be related to the emplacement of the external granitic plutons at about 2704 to 2699 Ma. Inset level plan and section of the former Madsen gold mine from Horwood (1945) show *en échelon* tension veins in the Austin shear zone, reflecting right-handed wrench movement.

in the batholith have been intruded by muscovite- and garnet-bearing pegmatite dikes from this body. A smaller satellite stock occurs nearby at the easternmost end of Lake St. Joseph.

Summary

The Uchi Subprovince has been the focus of several periods of volcanism and plutonism. The most prolific felsic plutonic activity appears to be late in the Kenoran Orogeny, about 2710 to 2690 Ma, represented by the massive granite to granodiorite bodies. But there are intrusions of trondhjemite to granodiorite composition that have been shown to have been emplaced at widely different times, prior to the Kenoran Orogeny. These earlier plutons may correspond to pre–2.7 Ga orogenic events in the Uchi Subprovince. Plutons of 2.7 Ga age, cutting older parts of the subprovince, display various forms and timing relative to orogenic activity. Most intriguing are the plutons in the northern half of the subprovince that are comparable in age to volcanism farther south, and yet locally show late to posttectonic relationships to the supracrustal rocks. Thus, the role of plutonic bodies as markers in the relative sequence of events in the crust can be controversial, particularly where the distinguishing criteria of relative timing based upon field relations are equivocal or not fully established. Since most felsic intrusions have only been examined in reconnaissance mapping, there remains considerable opportunity for enhancing our interpretation of the tectonic sequence of events. Several of the plutons in

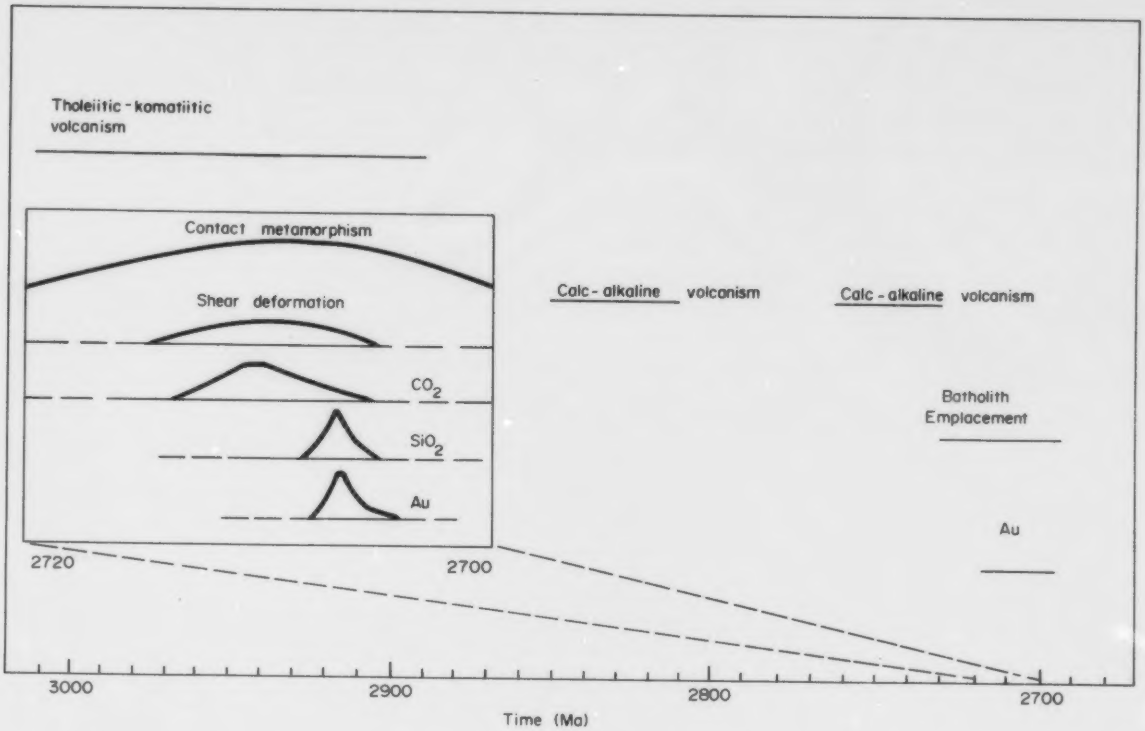


Figure 6.17b. A summary of the major events in the Red Lake greenstone belt from Andrews et al. (1986). Episodes of volcanism changed character through time and are followed by plutonic activity from about 2718 to 2699 Ma during which gold mineralization occurred. SiO_2 and CO_2 represent alteration associated with deformation zones; Au represents gold mineralization.

this subprovince have been used to constrain events summarized in Figures 6.17b and 6.18a to 6.18d.

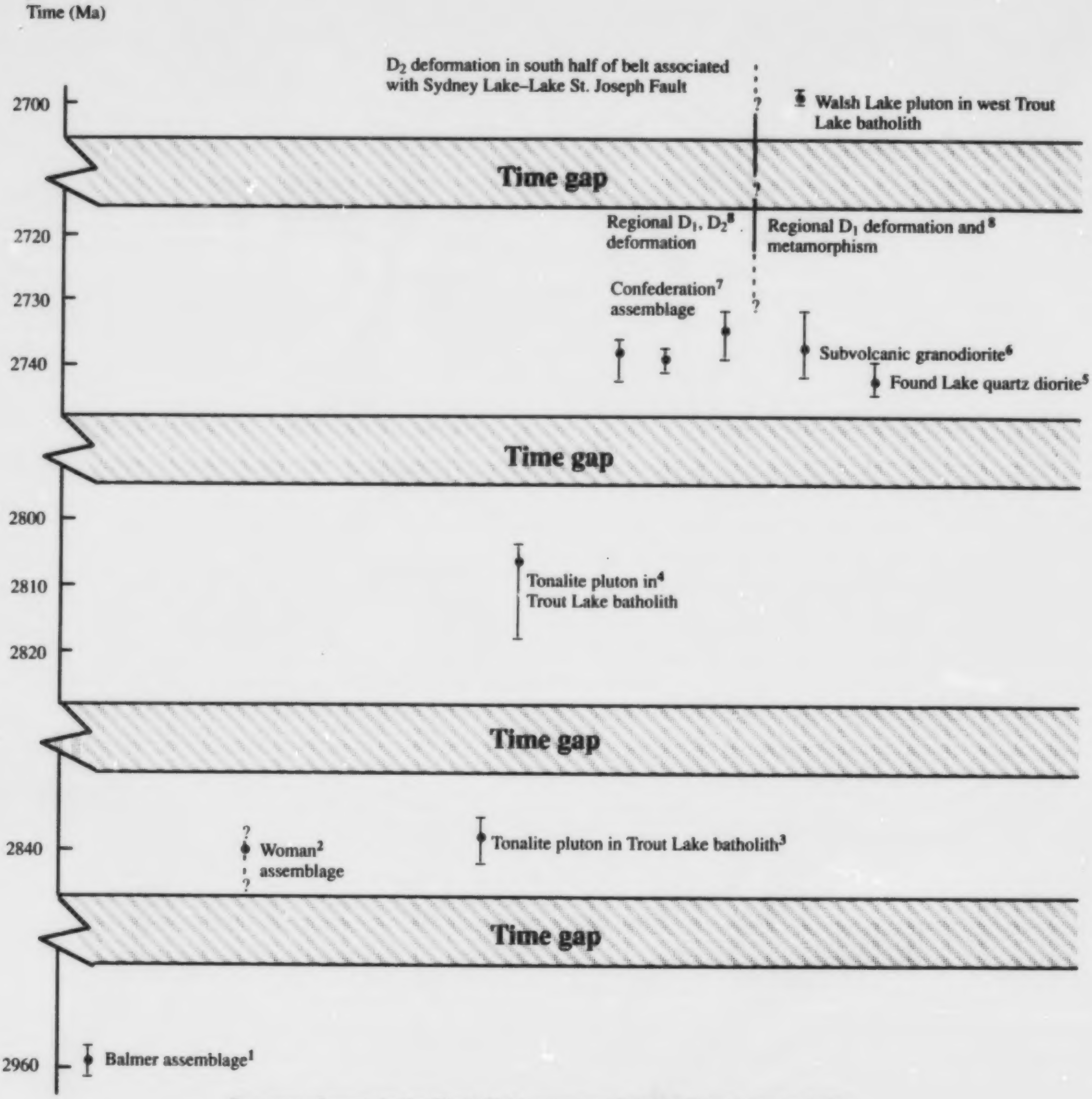
TECTONIC ELEMENTS—STRUCTURES AND METAMORPHISM

Structures

HISTORICAL BACKGROUND

The Uchi Subprovince and its boundaries have been subjected to limited structural analysis in selected areas, notably parts of the Red Lake greenstone belt where investigations were influenced by the presence of major gold discoveries and the search for more deposits. Horwood (1945) provided documentation of the structural relationships of ore bodies to the shear zones and offered an early structural analysis of the geology of some gold mines. The structural geology of some of the individual gold mines in Red Lake has been described in a symposium volume on Canadian ore deposits (CIMM 1948), and recently by Rigg and Helmstaedt (1981), Rigg and Scherkus (1983) and Sanborn (1987). In mapping the western Lake St. Joseph area in the early 1960s, Clifford (1969, 1972) became one of the earliest workers to provide a detailed demonstration of the value of lineations in defining, with fold patterns, the sequence of structural events in the greenstone belts. Recent reconnaissance and detailed studies of the deformation zones (defined in the chapter entitled Geology of Ontario: Introduction) were reported by Durocher and Burchell (1983), Durocher and Hugon (1983), Andrews et al. (1986), Hugon and Schwerdtner (1988), Wilson et al. (1984) and Berger (1984). This investigative activity arose from the increased gold exploration in the 1980s. Reconnaissance assessments of the structural domains of the greenstone belts have recently been conducted in the Birch–Uchi greenstone belt (Fyon and Lane 1985; Fyon and O'Donnell 1986) and in the Lake St. Joseph–Pickle Lake area (Stott and Wilson 1986a, 1986b; Stott et al. 1987a, 1987b; Stott et al. 1989a, 1989b).

An array of primary depositional structures in sedimentary and volcanic rocks, as well as magmatically and tectonically induced structures, provide us with patterns of stratigraphic younging, tectonic stacking and localized folding that characterize this granite-greenstone subprovince. This section will summarize the structural styles and regional patterns and summarize the structural elements and the local sequence of structures produced by regional, orogenic shortening and by intrusion of individual plutons into the major greenstone belts; these features provide us with the tools for erecting a general sequence of structural events across the subprovince as described in the section entitled Sequence of Events in the Uchi Subprovince.



Sequence of events in the Birch–Uchi greenstone belt and Trout Lake batholith

8 Regional deformation events.

7 Confederation assemblage volcanism (2739± 2; 2738$^{+5}_{-2}$; 2735$^{+4}_{-3}$ Ma).

6 Subvolcanic granodiorite (2737± 5 Ma).

5 Metamorphosed synvolcanic quartz diorite, Found Lake stock (2741$^{+3}_{-2}$ Ma).

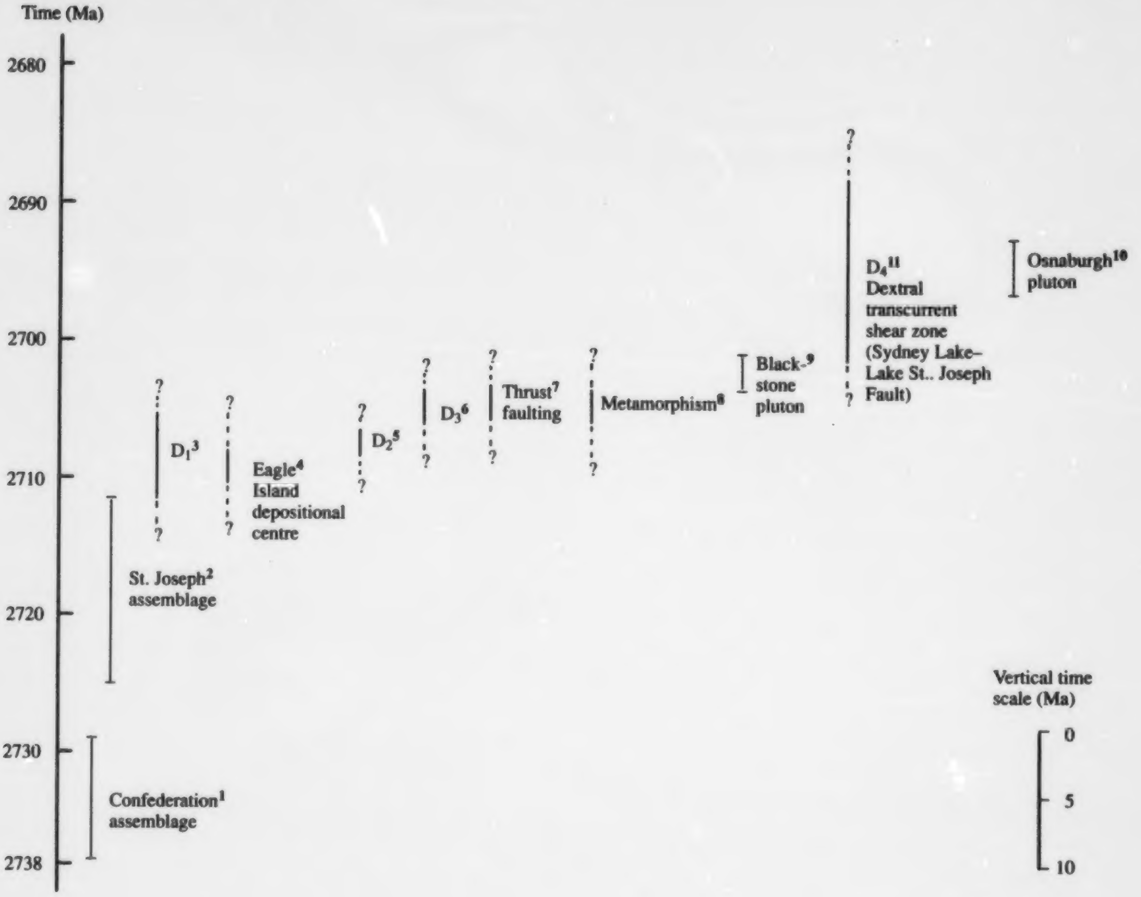
4 Tonalite (2806$^{+12}_{-2}$ Ma) stitching the boundary between Balmer and Woman assemblages.

3 Tonalite (2838$^{+4}_{-3}$ Ma) intruded margin of Balmer assemblage and is synchronous with Woman assemblage volcanics; developed on a continental margin.

2 Woman assemblage volcanism (ca. 2840 Ma), continental margin island arc, local stromatolite-bearing, 2840 million-year-old shallow-water carbonate; assemblage may have developed on a 2.96 billion-year-old North Caribou–Pickle composite terrane continental margin.

1 Balmer assemblage volcanism (2959± 2 Ma); oceanic mafic plain, tholeiitic to komatiitic.

Figure 6.18a. The general sequence of magmatic and structural events in several greenstone belts of the Uchi Subprovince. **a)** Birch–Uchi greenstone belt; **b)** Lake St. Joseph greenstone belt; **c)** Meen–Dempster greenstone belt; **d)** Pickle Lake greenstone belt. Published and unpublished ages by F. Corfu, S. Noble and P. Nunes, from the Jack Satterly Geochronology Laboratory, Royal Ontario Museum.



Sequence of events in the Lake St. Joseph greenstone belt

11 D_4 deformation: transpressive deformation within 1.3 km of subprovince boundary to produce dextral transcurrent shear zone with abundant kinematic indicators; followed by development of brittle, conjugate and asymmetric kink bands (D_3 of Clifford 1969).

10 Intrustion of crescent-shaped Carling pluton and Osnaburgh pluton (2693 ± 1 Ma).

9 Intrusion of granodioritic Blackstone pluton (2702 ± 1 Ma).

8 Metamorphism of Pembina Tonalite Gneiss of unknown duration (2704 ± 2 Ma, titanite).

7 Initial thrust faulting (Sydney Lake–Lake St. Joseph Fault) accompanying or subsequent to D_3 deformation.

6 D_3 deformation (D_2 of Clifford 1969): north-south shortening of Uchi–Sachigo microcontinent forming south-verging upright anticlinal fold.

5 D_2 deformation (D_1 of Clifford 1969): subsidence, slumping and infolding of Eagle Island depositional centre (restricted to depositional basin).

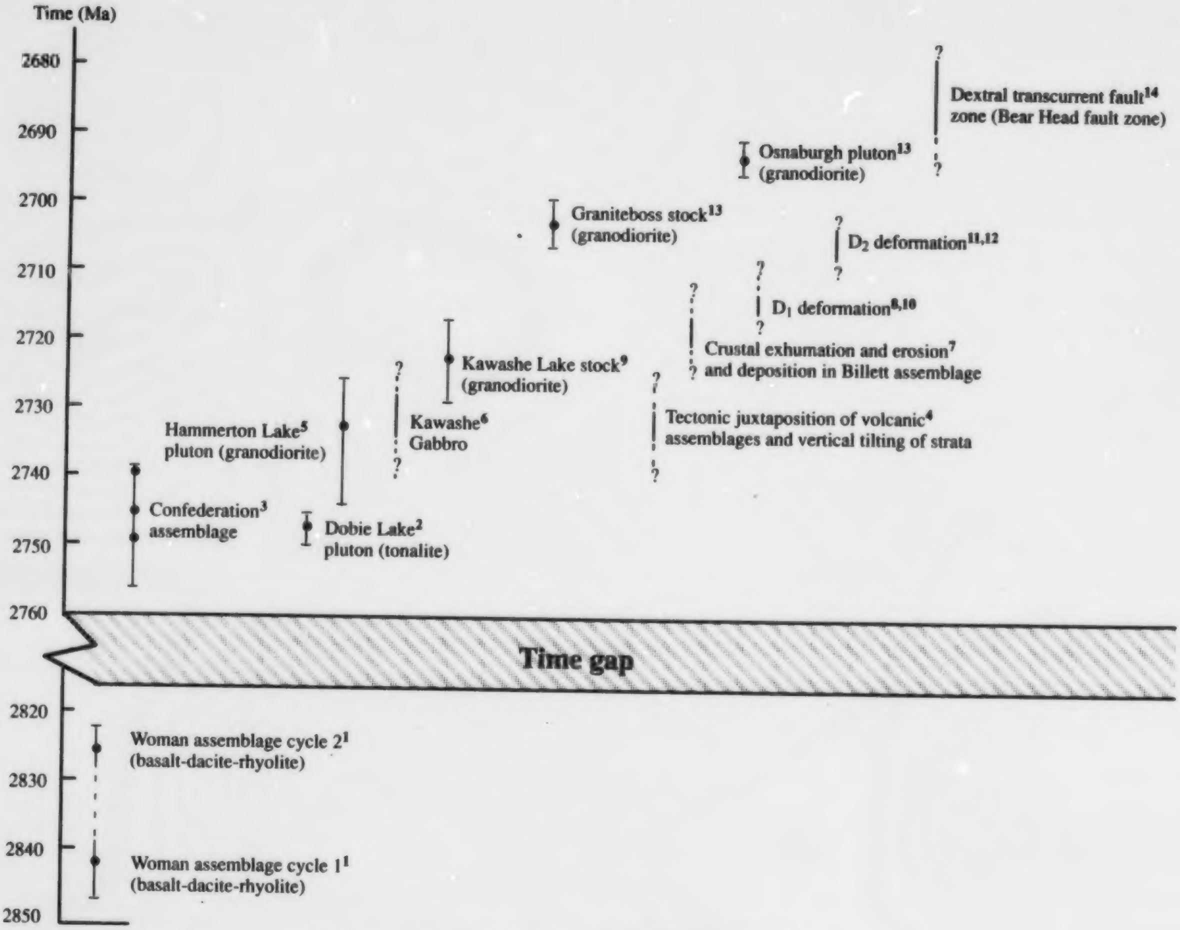
4 Crusal exhumation, erosion and deposition forming Eagle Island depositional centre; unconformably upon the east-facing upper felsic unit of the St. Joseph assem blage; basin accumulated wacke-arenite submarine fan delta and thick banded magnetite beds; basin may be proximal channel fill on margin of English River fore-arc basin.

3 D_1 deformation: tectonic stacking of Confederation and St. Joseph assemblages to form east-facing, steeply dipping sequences in west Lake St. Joseph.

2 St. Joseph assemblage volcanism; continental margin, island arc; tholeiitic basalt to calc-alkalic andesite-dacite-rhyolite.

1 Confederation assemblage volcanism; continental margin, island arc, bimodal tholeiitic basalt and rhyolite to dacite.

Figure 6.18b. The general sequence of magmatic and structural events in several greenstone belts of the Uchi Subprovince. **a)** Birch–Uchi greenstone belt; **b)** Lake St. Joseph greenstone belt; **c)** Meen–Dempster greenstone belt; **d)** Pickle Lake greenstone belt. Published and unpublished ages by F. Corfu, S. Noble and P. Nunes, from the Jack Satterly Geochronology Laboratory, Royal Ontario Museum.



Sequence of events in the Meen–Dempster greenstone belt

14 Bear Head fault zone – dextral transcurrent movement postdates plutons typical of Williams suite (2701 Ma, granite-granodiorite) and other plutons during the late stage of Kenoran orogenesis; fault may be reactivation of earlier fault in north Meen–Dempster greenstone belt margin.

13 Crescent-shaped Osnaburgh pluton (2693± 1 Ma) intruded between Meen–Dempster greenstone belt and foliated tonalite of Lake St. Joseph batholith; Graniteboss stock (2701 ± 2 Ma) postdates D_2 deformation.

12 D_2 deformation locally affected by *circa* 2700 Ma plutons.

11 D_2 deformation domain postdates Kawashe Lake stock and predates Graniteboss stock; D_2 possibly related to Kenoran orogenic northwest-southeast shortening producing translation and clockwise rotation of Dobie Lake pluton.

10 D_1 deformation fabric imposed on Billett assemblage.

9 Kawashe Lake stock intruded Kawashe Gabbro and was later subjected to D_2 deformation.

8 Dempster Gabbro intruded Billett assemblage and itself is deformed later by D_1 and D_2 deformation.

7 Crustal exhumation and erosion to produce Billett assemblage sedimentary deposition; interbedded conglomerate and graded sandstone; dominantly turbidite sequences unconformably overlie volcanic assemblages; some detritus evidently from erosion of Dobie Lake pluton (tonalite clast 2748^{+3}_{-4} Ma).

6 Kawashe Gabbro (ca. 2740 to 2722 Ma) stitches boundary between Woman and Confederation assemblages and itself is intruded by Kawashe Lake stock

5 Hammerton Lake pluton (2732^{+11}_{-7} Ma) intruded north margin of Meen–Dempster greenstone belt; pluton is synchronous with volcanism in south half of Uchi Subprovince

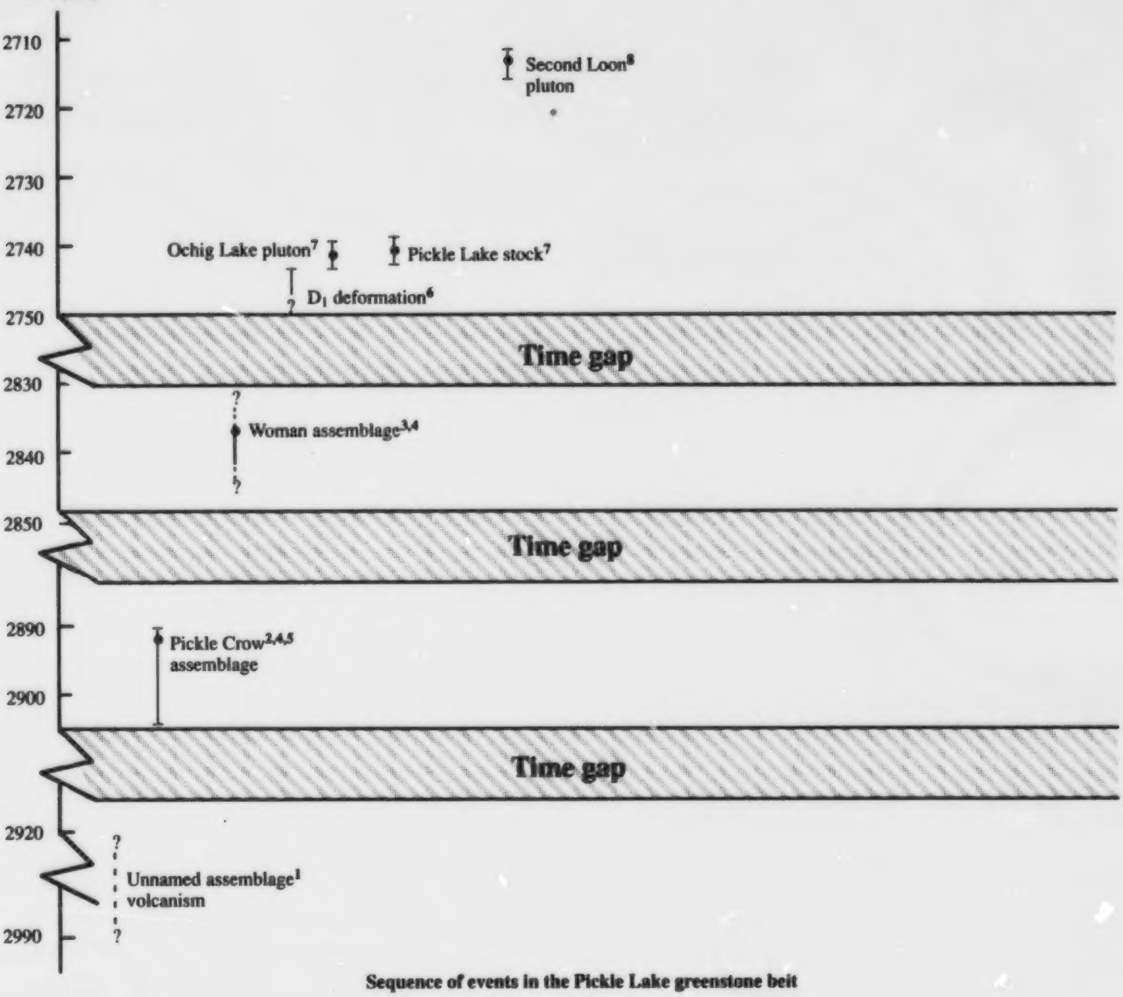
4 Local tectonic juxtaposition (by thrusting?) of Woman and Confederation assemblages bracketed between 2740 and 2722 Ma.

3 Confederation assemblage (2749 to 2740 Ma) volcanism; island arc or continental arc, bimodal tholeiitic basalt and calc-alkalic dacite.

2 Dobie Lake pluton (tonalite) intruded synchronously with Confederation volcanism; pluton imposed contact metamorphism on north margin of Meen–Dempster greenstone belt.

1 Woman assemblage (2841 to 2825 Ma) volcanism; continental margin island arc, bimodal tholeiitic basalt and calc-alkalic dacite to rhyolite.

Figure 6.18c. The general sequence of magmatic and structural events in several greenstone belts of the Uchi Subprovince. **a)** Birch–Uchi greenstone belt; **b)** Lake St. Joseph greenstone belt; **c)** Meen–Dempster greenstone belt; **d)** Pickle Lake greenstone belt. Published and unpublished ages by F. Corfu, S. Noble and P. Nunes, from the Jack Satterly Geochronology Laboratory, Royal Ontario Museum.



Sequence of events in the Pickle Lake greenstone belt

8 Granodioritic Second Loon pluton intruded (2713 ± 2 Ma) along greenstone belt margin and produced contact strain aureole (D_3 domain) in Pickle Lake greenstone belt.

7 Ochig Lake pluton (2741 ± 2 Ma), Pickle Lake (2740 ± 2 Ma) and Hooker–Burkoski, stocks intruded and produced contact strain and metamorphic aureoles, forming part of D_2 strain domain in Pickle Lake greenstone belt.

6 D_1 domain deformation postdates volcanic assemblages and predates 2741 million-year-old plutons.

5 Tectonic accretion of north half of belt with Pickle Crow assemblage evident from juxtaposed strata; northern assemblage strata, including iron formation and electromagnetic conductor marker-units, transected by Pickle Crow assemblage at a small angle. Accretion at undefined time.

4 Disconformable relation between Pickle Crow and Woman assemblages at undefined time.

3 Woman assemblage volcanism (ca. 2836 Ma).

2 Pickle Crow assemblage volcanism (2892^{+10}_{-2} Ma); island arc volcanism or rifted basin tholeiitic basalt and calc-alkalic rhyolite. Comparable in age to Bruce Channel volcanism in Red Lake greenstone belt.

1 Possible correlation of north half of Pickle Lake volcanism with McGruer assemblage in North Caribou greenstone belt (North Caribou terrane); oceanic mafic plain volcanism with tholeiitic basalt, ultramafic sills and iron formation.

Figure 6.18d. The general sequence of magmatic and structural events in several greenstone belts of the Uchi Subprovince. **a)** Birch–Uchi greenstone belt; **b)** Lake St. Joseph greenstone belt; **c)** Meen–Dempster greenstone belt; **d)** Pickle Lake greenstone belt. Published and unpublished ages by F. Corfu, S. Noble and P. Nunes, from the Jack Satterly Geochronology Laboratory, Royal Ontario Museum.

REGIONAL PATTERNS AND STRUCTURAL STYLES

The Uchi Subprovince is composed of irregularly shaped greenstone belts that weave around elliptical granitoid batholiths and also host smaller plutons. The complexity of this regional pattern, however, is tempered by the long, tabular shape of this subprovince. This tabular shape governs the dominant theme of easterly striking, asymmetric stratigraphic and structural patterns summarized here.

Stratigraphic Patterns

1. Tracing of assemblage subdivisions across the subprovince reveals a pattern of long tabular volcanic assemblages that approximately mirror the shape of the subprovince. The greenstone belts are, therefore, merely stratigraphic remnants engulfed in granitoid batholiths of widely varying age.
2. There is a general asymmetry of stratigraphic age across the breadth of the subprovince with age decreasing towards the south; there are exceptions to this pattern which may represent either later tectonic displacements of younger material towards the north (e.g., Confederation assemblage in northern Red Lake greenstone belt), or autochthonous ensialic rift basins of younger age within the northern half of the subprovince.
3. Plutons, which are synchronous with the episodes of volcanism, occur either close to coeval volcanic sequences (and intrude older assemblages in some cases), or they intruded farther north of the coeval assemblages. The presence of these plutons implies that either 1) most volcanic assemblages developed on the southern flank of a southward-growing continent, or 2) Archean lithospheric plates moved at a sufficient rate to transport volcanic terranes to an accreting continental margin, at the time of concurrent inboard plutonic activity. The first scenario is the simplest to accommodate.
4. There is a general geochemical asymmetry of volcanic rocks across the breadth of the subprovince. The oldest volcanic rocks, in the northern half of the subprovince, are dominated by tholeiitic and komatiitic basalt, whereas most volcanic assemblages comprise bimodal cycles of tholeiitic basalt and calc-alkalic dacite and rhyolite. The youngest volcanic cycles, in the southern half of the subprovince, display greater volumes of calc-alkalic dacite and some andesite.
5. There is a temporal progression in the petrographic nature of plutonic rocks. Tonalite is more voluminous amongst the older plutons and gneisses, and is succeeded in time by a greater volume of granodiorite and monzogranite plutons. The late to posttectonic plutons are represented across the breadth of the subprovince; late plutons in the northern part of the subprovince include a greater abundance of magnetite-bearing granite plutons, forming a chain parallel to the Uchi Subprovince, not unlike the Coastal Batholith of Peru in size and complexity (Pitcher 1985) if one includes the plutonic Bienville Subprovince of Quebec (Card and Ciesielski 1986). It is conceivable that this plutonic chain continues on another Archean shield.

Structural Styles

1. The northern half of the subprovince is marked by greenstone belts that trend to the northeast and northwest, which contrast with the dominantly eastward trend of the belts close to the southern boundary. This change across the breadth of the subprovince appears to reflect older strikes of bedding and schistosity in the northern half, similar to structures in the North Caribou terrane (*see* Thurston, Osmani et al., this volume), and younger folding and flattening during northward-directed shortening in the southern half.
2. The structural style across the Uchi Subprovince is characterized by moderately deformed zones of volcanic strata bordered by narrow displacement zones of more intense strain. Folds are concentrated either near plutons or in the southern half of the subprovince. Most greenstone belt sequences are steeply dipping; steep dips of bedding and schistosity occur commonly in a single direction across broad parts of the belts or across the entire width of a belt. For example, the bedding and schistosity dip northward in the Meen–Dempster greenstone belt and major parts of the Lake St. Joseph greenstone belt.
3. The degree of penetrative strain in these rocks is typically moderate to weak, and consequently, substantial parts of the greenstone belts display an excellent preservation of primary volcanic and sedimentary structures as illustrated in Figures 6.19a, 6.19b and 6.19c. But not all areas took up their share of the strains imposed by regional, tectonically induced compression or pluton emplacement. Consequently, linear zones of ductile to brittle-ductile deformation (often referred to as regional-scale shear zones or faults) can be traced across virtually all greenstone belts and examples can be seen in Figures 6.4 and 6.9.
4. Given the belt-like form of the volcanic assemblages, it is not surprising that the structural style of large parts of greenstone belts in the Uchi Subprovince is dominated by tabular zones of volcanic strata that consistently face in a single direction (referred to here as "homofacing" strata). These zones comprise panels, several kilometres thick, of volcanic flows and pyroclastic rocks with limited sedimentary units. Such panels of "homofacing" strata typically consist of one or more assemblages. An example of this is the northern part of the Woman assemblage, which is eastward facing in the Birch–Uchi greenstone belt and becomes southward facing in the Meen–Dempster and Pickle Lake greenstone belts; it is overlain by the similarly eastward- and southward-facing Confederation assemblage (*see* Figure 6.3). In some cases where U-Pb isotopic ages show that the strata are outside of the normal order of superposition, the geochronology serves as indirect evidence of tectonic juxtaposition by faulting even though the

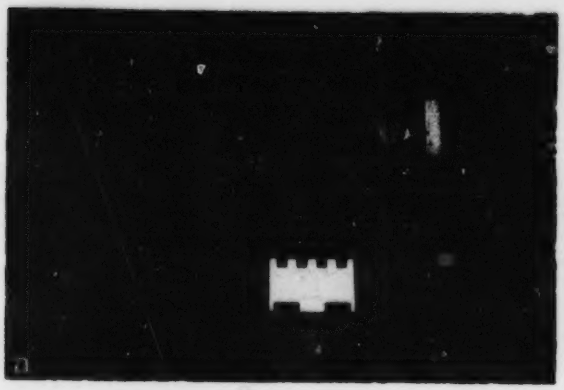


Figure 6.19a. Well-preserved, bun-shaped basaltic pillows with radiating cooling fractures in the Pickle Lake greenstone belt. Pillow-packing is the most widely used stratigraphic top indicator.

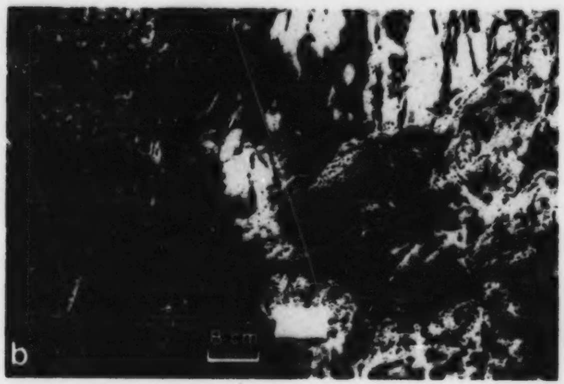


Figure 6.19b. A basaltic sequence of flows in the Confederation assemblage of western Lake St. Joseph (cycle 3), showing well-defined contact between a massive basaltic flow and an underlying pillowed flow. The schistosity is orthogonal to bedding since this outcrop is on the axial plane of the major anticline in the southern part of the Lake St. Joseph greenstone belt. Top directions in this volcanic cycle are eastward, facing away from the synclinal Eagle Island assemblage and indicate that the latter assemblage is unconformably overlying the volcanic assemblages (*see also* Figure 6.6).

fault has not yet been directly observed. This can be seen from the relative stratigraphic positions of geochronology sites in the Birch–Uchi and Meen–Dempster greenstone belts (*see* Figures 6.5 and 6.9).

These thick "homofacing" strata may, therefore, be bounded by disconformities and faults that closely parallel the stratigraphic layering. The panels of strata can be inferred from Figure 6.3; they may comprise a simple stratigraphic succession or a tectonically thrust-stacked sequence of either repeated strata or strata out of normal sequence. Adjacent stratigraphic panels that face towards each other typically mark the presence of a faulted assemblage boundary, for example, within the deformation zones of the Red Lake greenstone belt (*see* Figure 6.4). However, in some cases multi-kilometre-thick "homofacing" panels *within* an assemblage, face toward each other across an intra-assemblage fault. This is evident in the central part of the Confederation assemblage of the Birch–Uchi greenstone belt, where a syncline was formerly interpreted (Thurston 1985b).

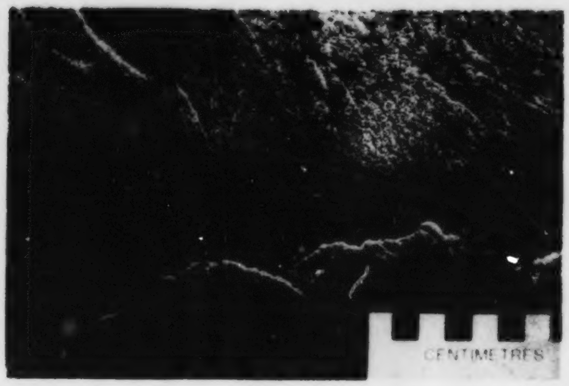


Figure 6.19c. Flame structures in arenaceous sandstone beds of the Eagle Island assemblage. Such structures are well preserved in this assemblage but are locally asymmetrical, having rotated towards the axial planar cleavage direction of a fold.

5. Most of the greenstone belts contain an older fabric domain with obliquely plunging, mineral-elongation lineations and shape-fabric lineations, which are subparallel to the strike of this subprovince and resemble orogen-parallel lineations formed in Phanerozoic orogens (Ellis and Watkinson 1987). Closer to the southern boundary and in the strain aureoles and shear zones around plutons, there is greater evidence for downdip plunge of lineations.
6. Folds and shear zones are concentrated near plutons and near the southern boundary of the subprovince. Folds near the Uchi–English River subprovince boundary are tight, upright and trend subparallel to the boundary. In the Birch–Uchi and the Miminiska–Fort Hope greenstone belts, east-striking shear zones increase in prominence as the southern margin of the Uchi Subprovince is approached; this spatial relationship and similar dextral kinematics suggest these shear zones overprint the early folding and are associated with the ductile, transcurrent, Sydney Lake–Lake St. Joseph fault zone along the subprovince boundary.
7. The regional-scale faults strike dominantly eastward and southeastward with evidence of consistent right-handed strike-slip displacement. The orientations of these faults and their sense of displacement are consistent with a north-northwest-directed shortening of the crust in this region.

In summary, folds and accompanying shear zones in volcanic and sedimentary strata appear to be preferentially concentrated near external plutons and near the boundary with the English River Subprovince (*see* Figure 6.6); the larger greenstone belts are otherwise predominantly characterized by broad "homofacing" panels of strata bounded by faults and shear zones.

The sequence of structures and the geochronology of plutons and volcanic units indicate several ages of deformation. The evidence shows that, although the structures related to the main phase of the Kenoran Orogeny dominate the southern half of the Uchi Subprovince, there are older structural domains, in the north half of the subprovince (e.g., the Pickle Lake greenstone belt), which indicate similar sequences of events. However, the best preserved sequence of events is based on structures related to the Kenoran Orogeny. The timing of structural events is bracketed by the emplacement of plutons as summarized for the major belts in Figures 6.18a to 6.18d. Indirect evidence from the interleaving of volcanic assemblages in some belts implies that part of the Kenoran Orogeny involved an initial stage of thrust faulting, with tectonic displacement of strata northward, away from the Uchi–English River boundary. The volcanic and sedimentary strata were tilted steeply and stacked as thick "homofacing" panels. Regional-scale penetrative shortening followed to produce orogen-parallel stretching lineations in the interior of the larger greenstone belts and dip-slip stretching in the strain aureoles of plutons and in the vicinity of the southern boundary of the subprovince. The regional ductile shortening produced folds, typically adjacent to shear zones as in the Birch Lake area and the Pickle Lake greenstone belt. This Kenoran orogenic shortening was accompanied and followed by plutons rising and in some cases squeezing the surrounding wall rocks as the plutons intruded them. The terminating stage of deformation in the Uchi Subprovince is marked by transcurrent faults that extend for 100s of kilometres and involved displacement and rotation of crustal blocks. The orientations of folds, faults and accompanying shear folds near the southern boundary, indicates that the principal direction of shortening during the Kenoran Orogeny was along a north-northwest axis. The style of structures during this orogeny progressed from 1) early thrust faulting, through 2) ductile flattening and folding on a regional scale and locally induced by plutons, and concluding with 3) ductile and later brittle transcurrent faulting.

The following sections, intended to highlight regional structural patterns in each of the main greenstone belts, provide some illustrations of these features.

REGIONAL PENETRATIVE STRUCTURES

Schistosity, Lineations and Folds

There is sufficient regional-scale mapping of structures in the greenstone belts to permit a summary of the general patterns of planar and linear fabric elements and folds.

SCHISTOSITY

In most areas, a single planar alignment of minerals (schistosity) is observed to dip steeply or vertically and there may be a dominant direction of the steep dip, for example, the Meen–Dempster greenstone belt, where the schistosity dips vertically or steeply to the north. The schistosity typically lies either parallel to or at a low angle to the bedding and strikes subparallel to the length of the greenstone belt or conforms to the boundaries with external batholiths. The schistosity is, therefore, potentially attributable to either regional orogenic shortening along the length of the subprovince or to buoyant or forceful emplacement of the granitic intrusions. Pronounced contrast in flattening strain intensity can be observed in detail from the varioles, pillow selvages and other strain markers in all of the greenstone belts. On a regional scale, the intensity of schistosity development varies such that one can map sizeable areas where the flattening fabric is weak and contrasts with narrow zones of intense ductile and brittle-ductile deformation. Typically, these zones of deformation are localized along older faults or closely follow certain stratigraphic units, such as iron formation layers and quartz porphyry units that contrast in ductility with the surrounding rocks.

LINEATIONS

Variations in trend and plunge of lineations have been used to define structural domains (e.g., in the Pickle Lake greenstone belt) that can be attributed to either regional shortening or to the emplacement of local plutons. Domains produced or modified by pluton emplacement are discussed in the section entitled Pluton-induced Structures. Lineations, consisting of mineral elongations and shapes of mineral aggregates, not directly related to the emplacement of granitic plutons, plunge moderately to steeply subparallel to the eastward trend of the subprovince; that is, the lineations are orogen-parallel. These are most evident in domains of the northern half of the subprovince (Stott and Wilson 1986b; Stott et al. 1987a); but such domains are less well preserved and thus less prominent in some belts where broad strain domains were produced by the emplacement of particular granitic plutons as in the northeastern part of the Birch–Uchi greenstone belt (Good 1988). Closer to the Uchi–English River boundary, lineations typically plunge steeply downdip on the schistosity plane, for example, in the Lake St. Joseph greenstone belt (Clifford 1969). But within several hundred metres of this southern boundary, in the Sydney Lake–Lake St. Joseph fault zone, the lineations plunge shallowly to moderately (0° to 30°) eastward; this lineation orientation is parallel to the predominant lineation and fold axis orientation in the English River Subprovince (*see* Breaks, this volume) and indeed, it is parallel to the D_2 lineations and fold axes in the Winnipeg River (*see* Beakhouse, this volume) and northern Wabigoon subprovinces (*see* Blackburn et al., this volume) and in the Quetico (*see* Williams, this volume) and northern Wawa subprovinces (*see* Williams, Stott, Heather et al., this volume). This distinctive, eastward-plunging linear fabric characterizes broad zones near the southern margin of the Uchi Subprovince and across the adjacent English River metasedimentary subprovince (*see* Breaks, this volume). This fabric is interpreted to reflect orogen-parallel displacement of ductile rock during late Kenoran transpression across the Superior Province (see Williams, Stott, Thurston et al., this volume)

The boundaries between structural domains are in some cases narrow zones of transition in lineation orientation (e.g., in northwestern Meen–Dempster greenstone belt, Figure 6.20). In other cases they are shear zones (e.g., between domains D_1 and D_2 of the Meen–Dempster greenstone belt, Figure 6.9). In most domains, there is only one major mineral lineation in the volcanic rocks with rare evidence of overprinting by a second lineation. Consequently, the volcanic domains are less revealing about the structural history than sedimentary domains which exhibit several generations of folds and lineations; one need only contrast the detailed record of superimposed planar and linear fabric elements preserved in the sedimentary rocks, for example, the Eagle Island assemblage of Lake St. Joseph documented by Clifford (1969), with the less complex record preserved in the volcanic rocks.

FOLDS

The earliest folds in the northern part of the subprovince are tight, layer-parallel with sharp changes in stratigraphic younging directions (e.g., Birch–Uchi greenstone belt, Beakhouse 1989; Pickle Lake greenstone belt, Thomson 1939a; Bamaji Lake area of the Meen–Dempster greenstone belt, Wallace 1985). Later, D_2 folds, which buckle the D_1 flattening fabrics, vary from open to tight and are attributed to either the emplacement of adjacent external granitic batholiths (as in the northern part of the Birch–Uchi greenstone belt; Good 1988; Beakhouse 1989) or, to regional orogenic shortening, especially near the southern subprovince boundary. The D_2 event includes increased deformation intensity, evident near the southern subprovince boundary; this folding is spatially and temporally related to late shear zones (as in the southern part of the Birch–Uchi greenstone belt; Fyon and Lane 1985; Williams 1988). This deformation is also visible as tight asymmetric folds observed near the southern boundary of the subprovince and within the Sydney Lake–Lake St. Joseph fault zone (e.g., the Bee Lake belt; Stone 1977, 1981).

The western part of Lake St. Joseph is characterized by a more open fold, with axial planar cleavage striking subparallel to the subprovince boundary; its shape is controlled largely by the original shape of the volcanic edifice that dominates that area (Clifford 1969).

In general, 2 major sets of folds can be observed in most belts; the sequence of folding typically begins with steeply plunging, layer-parallel, tight D_1 folds, refolded by open to tight D_2 folds. The D_2 folds may be either steeply plunging

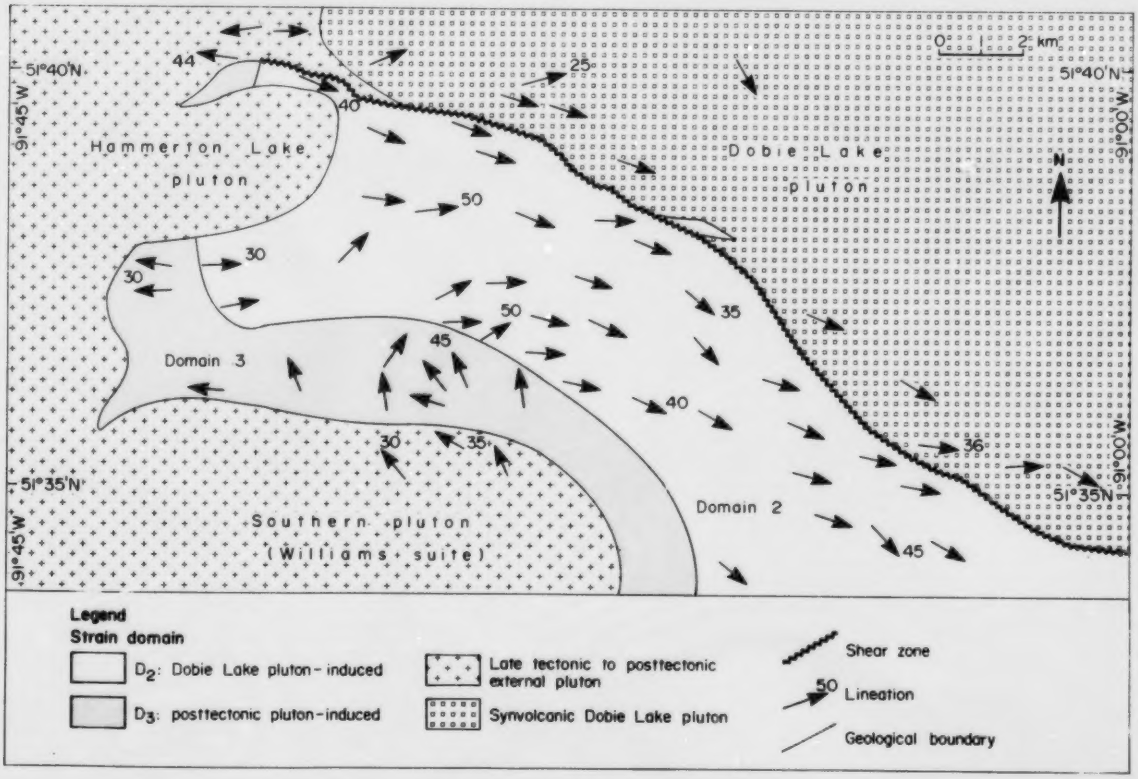

51°40'N
91°45'W
44
40
25
0 2 km
51°40'N
91°00'W
Hammerton Lake pluton
Dobie Lake pluton
N
50
30
30
35
Domain 3
50
45
40
36
30
35
51°35'N
91°45'W
Southern pluton (Williams suite)
Domain 2
45
51°35'N
91°00'W
Legend
Strain domain
D2: Dobie Lake pluton-induced
D3: posttectonic pluton-induced
Late tectonic to posttectonic external pluton
Synvolcanic Dobie Lake pluton
Shear zone
50
Lineation
Geological boundary

Figure 6.20. A map of mineral elongation lineations and shape ellipsoids of clasts and varioles in the northwestern end of the Meen–Dempster greenstone belt. The pattern of lineations correspond to 2 domains of strain in this area; the D_2 domain (*see also* Figure 6.9) shows consistent, moderately plunging lineations and includes the southern part of the adjacent Dobie Lake pluton; the D_2 domain is related to regional orogenic shortening and was overprinted along the southern margin of the belt by a contact strain aureole imposed by a late tectonic to posttectonic granodiorite pluton. D_2 lineations rotated or reformed within the schistosity to acquire the local D_3-northwestward plunge corresponding to lineations of the pluton's magmatic flow fabric. Measurements are from Stott and LaRocque (1983a).

and related to external batholith emplacement (e.g., in the northwestern Birch–Uchi greenstone belt, Beakhouse 1989), or moderately and shallowly plunging in response to regional northward-directed, subhorizontal shortening, which is most intense near the subprovince boundary (e.g., in the southern part of the Birch–Uchi greenstone belt, Williams 1988). Local D_3 folds are typically small, open folds, typified by crenulations and kinks and are most common near fault zones, for example, in the Birch–Uchi greenstone belt (Beakhouse 1989) and along the Sydney Lake–Lake St. Joseph fault zone (Stott et al. 1987a; Williams 1988).

A local sequence of deformation events can be determined from the relationships amongst the schistosities, lineations, folds, plutons and deformation zones as illustrated in the Meen–Dempster and Pickle Lake greenstone belts (*see* Figures 6.9 and 6.10). The following descriptions highlight some typical structural features in several of the greenstone belts of the Uchi Subprovince. These descriptions are not intended to be a subprovincial scale structural overview, rather, they are offered to emphasize particular features of the structural style and history of the greenstone belt in question. In some cases, new structural interpretations of a greenstone belt are offered, and in other cases a summary of the sequence of structural events in a greenstone belt is given.

Bee Lake Belt

The Bee Lake belt is a continuation of the Rice Lake greenstone belt of Manitoba; the pattern of the major units is a reflection of what is interpreted here as a fold geometry, the evidence for which is summarized in the following paragraphs. The southern part of the belt also illustrates some typical structures of a wrench fault zone. The Bee Lake belt is a southeastward-plunging, generally northward-dipping antiformal syncline, cored by felsic to intermediate pyroclastic rocks that correlate with the Banksian Lake Formation (Weber 1971a) in Manitoba. The structural facing direction (the direction towards younger strata) based on few top determinations in Ontario (Shklanka 1967) and Manitoba (Poulsen 1989), is northwestward along the antiform axis. Farther southeastward, the clastic sedimentary rocks along the hinge of the antiform structurally overlie the pyroclastic rocks, yet show tectonic interleaving with these rocks through the formation of secondary layering on a regional scale in the nose of the anticline. These sedimentary rocks are tectonically transposed along the limbs of the antiform and can be discontinuously traced to sedimentary rocks of the Rathall Lake Formation (Gem Lake Subgroup) in Manitoba (Weber 1971b). In similar style, massive to pillowed basalts structurally overlie these sedimentary rocks, forming secondary layers interleaved with the sedimentary rocks in the fold nose, and yet thicken to a 3.5 km section along the southern flank of the antiform. North of the belt, the basalt is represented by amphibolitic layers in gneissic granodiorite that can be traced into Manitoba (Weber 1971b). Such an eastward-plunging antiform, seemingly located on the stratigraphically thicker, felsic pyroclastic lenses (apex of a volcanic edifice?), is similar to the northward-dipping, easterly plunging anticline in the Lake St. Joseph greenstone belt described later in the chapter. In both cases, the major fold lies close to and parallel with the Uchi–English River subprovince boundary.

Lineations of minerals and stretched pyroclastic fragments of the Bee Lake belt typically plunge 30° to 70° westward and may reflect direction of shear during folding. This lineation orientation is regional in scope and can be traced throughout the Gem Lake Subgroup into Manitoba (McRitchie and Weber 1971b). Fold noses in the Rice Lake greenstone belt and its extension into Ontario are characterized by closures to the east and typically show megascale Z asymmetry, consistent with subhorizontal regional shortening by transpression along a northwest axis, oblique to the strike of the subprovince boundary. This eastward closing of fold noses also can be seen where English River sedimentary units interfinger at the boundary with the Uchi volcanic units and, on smaller scales, between units within the belt (Figure 6.21). A folded sequence of English River wacke deposits and amphibolitic basalt flows in the Bee Lake belt occurs in the vicinity of Chase Lakes (*see* Figure 6.21), and is partly separated from the main mass of English River metasedimentary rocks by a thin sequence of felsic tuff. This felsic tuff marks both the southern extent of the Uchi Subprovince and the presence of the Sydney Lake–Lake St. Joseph fault zone (Stone 1981). Asymmetric megascale Z folds occur on various scales close to this fault zone and are part of the fabric associated with transpressive deformation, observed near subprovince boundaries such as the English River–Winnipeg River boundary (Sanborn-Barrie 1988).

Red Lake Greenstone Belt

There is no published analysis of the regional structural framework of the Red Lake greenstone belt. Nevertheless, age determinations, stratigraphic mapping, local younging directions of bedding, and the identification of fault zones within the Red Lake greenstone belt (*see* Figure 6.4) provide the principal evidence for the following summary and a general sequence of events (*see* Figure 6.17b). The Red Lake greenstone belt is a collage of tectono-stratigraphic assemblages, juxtaposed along fault boundaries that lie within the deformation zones described by Andrews et al. (1986) and shown in Figure 6.4. Most of the assemblages are composed of steeply dipping, multi-kilometre-thick panels of strata, with each panel facing in a consistent direction. Strata dip subvertically or verge toward the external batholiths. The schistosity in the belt is parallel to the contacts with the external batholiths and contains mineral elongation lineations and shape lineations that plunge downdip, probably reflecting strain imposed upon the belt by adjacent, external plutons. These structures are overprinted by discrete shear zones in the deformation zones. Structural investigations in the Red Lake greenstone belt have focussed on the deformation zones, their relation to gold mineralization and the relative timing of gold to other magmatic and tectonic events as described by Andrews et al. (1986), Wilson et al. (1984) and Hugon and Schwerdtner (1988).

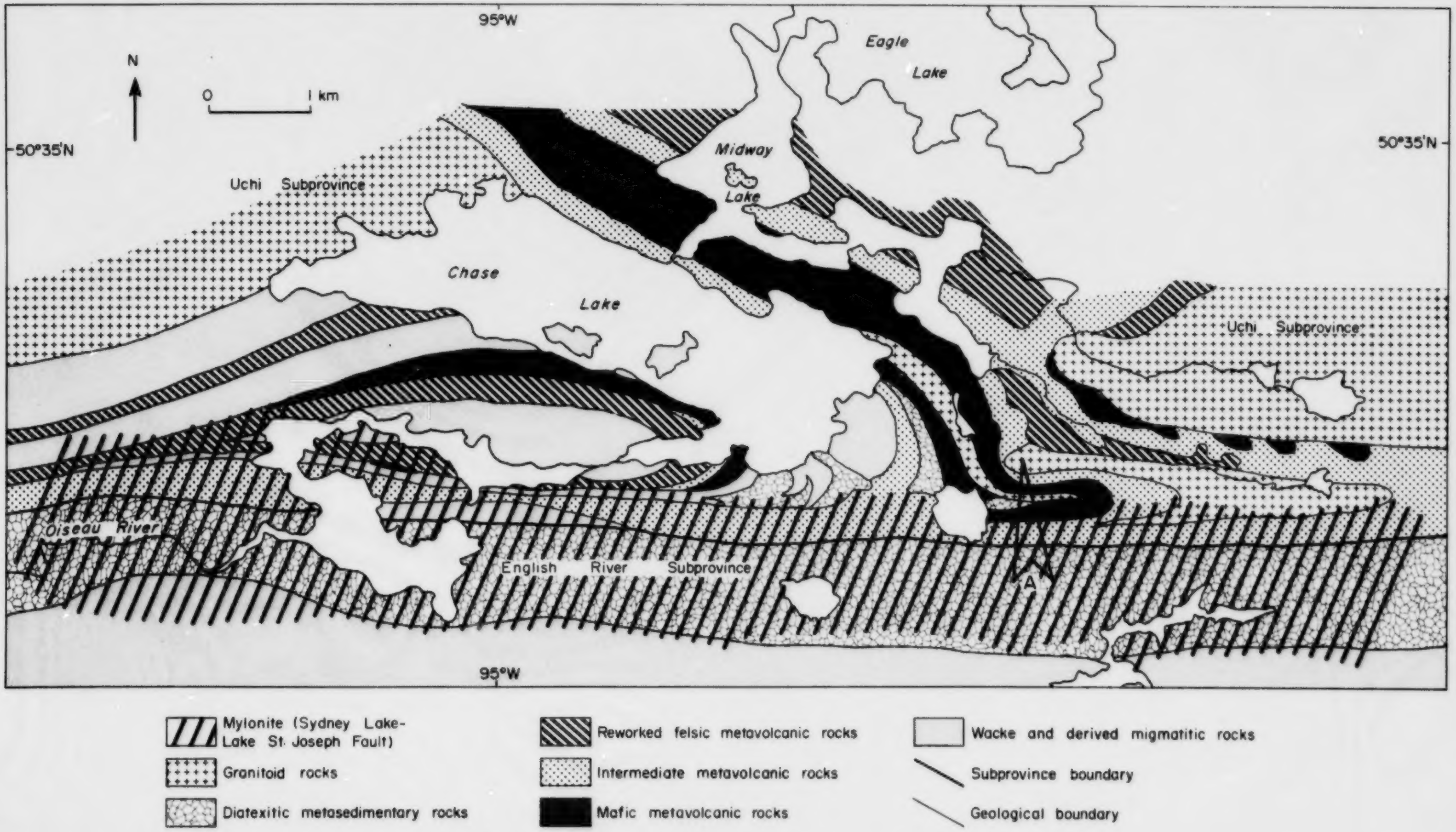

95°W
N
0
1 km
50°35'N
50°35'N
Eagle Lake
Midway Lake
Chase Lake
Uchi Subprovince
Uchi Subprovince
Oiseau River
English River Subprovince
A
95°W
Mylonite (Sydney Lake-Lake St. Joseph Fault)
Granitoid rocks
Diatexitic metasedimentary rocks
Reworked felsic metavolcanic rocks
Intermediate metavolcanic rocks
Mafic metavolcanic rocks
Wacke and derived migmatitic rocks
Subprovince boundary
Geological boundary

Figure 6.21. The Sydney Lake–Lake St. Joseph Fault separates the Uchi and English River subprovinces. Metavolcanic units of the Bee Lake belt, interlayered with granitic rocks, are folded and attenuated within the fault zone. Stone (1981) has shown from the folding of the metavolcanic units in this fault zone that a minimum right lateral displacement of 6 km occurred if the metavolcanic units originally extended southeast of point A and were subjected to simple shear (*modified from* Breaks et al. 1978).

A characteristic lack of large-scale folding is evident in the volcanic assemblages. However, large-scale, tight folding occurs in the eastern part of the belt, within the Bruce Channel assemblage and is traced particularly by magnetite-iron formation layers in the upper sedimentary units of this assemblage. The tectonic boundary between the Bruce Channel assemblage and the overlying, older Balmer assemblage is likewise folded. This fold pattern conforms to the margin of the Trout Lake batholith and is attributed to buckling during emplacement of the large, crescentic Walsh Lake pluton within the batholith.

In the southern half of the belt, the Confederation assemblage shows a change in younging direction across a long, narrow, late-tectonic granitic intrusion, which lies on strike with a layer-parallel fault in the Birch–Uchi greenstone belt. The granitic intrusion, similar to some other sheet-like intrusions along the Uchi–English River and Winnipeg River–Wabigoon subprovince boundaries, may have originated by fault-induced granitic magmatism as described by Leake (1990) for Hercynian fault-associated granite intrusions in Europe.

Faults are most evident between the Balmer and Ball assemblages, and between the Confederation assemblage and other, adjacent assemblages. At Trout Bay of Red Lake, the fault boundary between the Balmer and Ball assemblages separates oppositely facing volcanic strata of contrasting composition and age. It is uncertain when these two assemblages were initially juxtaposed. The boundary between the Confederation and Balmer assemblages is marked by the Flat Lake–Howey Bay deformation zone, which includes the Austin shear zone (*see* Figure 6.17a). This is a critical boundary in the sequence of events since: 1) it is transected by the 2718 million-year-old Dome stock (Corfu and Wallace 1986); 2) it was subsequently reactivated to form the deformation zone; and 3) it was thermally metamorphosed (*see* Figure 6.17a) during emplacement of the 2704 million-year-old Killala–Baird batholith (Corfu and Andrews 1987). Thus, the collage of assemblages in the Red Lake greenstone belt was constructed by 2718 Ma, before eruption of the rocks forming the 2713 million-year-old St. Joseph assemblage in the Lake St. Joseph greenstone belt.

Birch–Uchi Greenstone Belt

The Birch–Uchi greenstone belt is unique; its tectono-stratigraphic units strike dominantly northward in contrast to most other Uchi Subprovince greenstone belts. The eastern half of the belt possesses a curviplanar strike of bedding and schistosity and an array of 2.73 to 2.74 Ga ages (Noble et al. 1989) on volcanic strata within the same sector of the belt. These relationships led the authors to speculate that the belt's distinctive orientation is a consequence of northward tectonic transport, that is, allochthonous or parautochthonous displacement of Confederation assemblage strata during the Kenoran Orogeny prior to the emplacement of late-tectonic external plutons. It has yet to be demonstrated that the sequence of structures described below relate to such a displacement event.

In the southern part of the Birch–Uchi greenstone belt, a sequence of fabrics recording 3 deformation events has been documented by Fyon and Lane (1985) and Fyon and O'Donnell (1986). These observations are complemented by those of Beakhouse (1989) and Good (1988) in the northern part of the belt where in addition to a D_1 deformation, broad contact strain aureoles of D_2 fabrics have been imposed subsequently upon the belt by late external batholiths. The belt is characterized by a D_1 deformation consisting of a northward-striking schistosity, generally conformable to the bedding and conformable to the northward trend of the belt. This is accompanied by a subvertical stretching lineation, seen from the long axes of shape indicators such as quartz aggregates, clasts and varioles. The stretching lineation is typically parallel to the long axes of amphibole minerals and streaks of chlorite and mica. Layer-parallel axial planes as well as D_1 folds, have been reported by the above workers in the northern and southern parts of the belt. In the Birch Lake area for example, Beakhouse (1989) noted that D_1 folds are principally recognized through refolding by the regionally dominant D_2 folds. The significance of the D_1 folds is not known. The D_1 deformation may be a consequence of compression by external batholiths or perhaps regional deformation during allochthonous or parautochthonous northward transport and rotation of the belt strata towards a northerly strike. One might speculate that this deformation event may have accompanied the aggregation of the assemblages after 2730 Ma, during collision between the Uchi–Sachigo and the Wabigoon–Winnipeg River superterranes. This northward transport is conceived to have been in the form of a broad "fist-like", northward-closing protrusion of the assemblage, which overrode part of the Woman assemblage. Similarly, in the Red Lake greenstone belt, general northward displacement of the Confederation assemblage may have also caused part of the assemblage to become separated and transported to the northern flank of the belt (*see* Figure 6.3).

Conceivably associated with this northward transportation event in the Birch–Uchi greenstone belt is the second deformation (D_2), which becomes more prominent in the southernmost part of the belt as one approaches the English River Subprovince. However, it does not become fully penetrative and consequently coexists with the northward-striking D_1 fabric (Fyon and Lane 1985), which can be traced to the southernmost part of the belt. Near the faulted boundary with the English River Subprovince, Williams (1988) described D_1 and D_2 sets of deformation fabrics. The D_2 schistosity strikes eastward and dips moderately to steeply southward, carrying a shallowly plunging stretching lineation which is generally masked by the steeper D_1 lineation. Superimposed on the D_2 fabrics are locally developed crenulations related to the D_2 event or reflecting a subsequent deformation. In the Birch Lake area farther north, Beakhouse (1989) proposed that the second deformation in the Birch Lake region is a consequence of pluton emplacement into the external Trout Lake batholith. This D_2 deformation apparently bears no relationship to the D_2 structures documented by Fyon and co-workers in the southern part of the belt, where D_2 deformation increases in prominence southward, notably as shear zones, as one proceeds towards the subprovince boundary. The latter D_2 event may be attributed to late-stage orogenic shortening, similar to the prominent D_2 structures near the Sydney Lake–Lake St. Joseph Fault; these structures are described below.

In the central part of the belt, both D_1 and D_2 shear zones have been documented by Fyon and O'Donnell (1986). The D_2 shear zones strike northeastward obliquely across D_1 schistosity; the D_1 deformation zones follow the schistosity and are attributed to irrotational strain. Kinematic evidence for rotational strain in bona fide shear zones is not always forthcoming and some deformation zones are attributable to strong flattening in irrotational strain domains. Examples of these contrasting high strain zones have been described by Fyon and Lane (1985) and Fyon and O'Donnell (1986).

The style of the D_2 deformation near the subprovince boundary is similar to that documented across the Wabigoon–Winnipeg River subprovince boundary (Sanborn-Barrie 1988) and across the Quetico–Wawa subprovince boundary (*see* Williams, Stott, Heather et al., this volume).

Notably, both the eastern part of the Birch–Uchi greenstone belt and the southern Red Lake greenstone belt, comprising Confederation assemblage rocks, form large closures toward the north that in map view resemble allochthons verging northward. The northward direction of closure of these rocks compares with the northward apical forms of the Handy, Boyer and Kakagi assemblages in the Wabigoon Subprovince (*see* Blackburn et al., this volume). Consideration of these assemblages from this perspective has not been documented although Thurston and Breaks (1978) for different reasons had postulated a northward-verging nappe encompassing the entire Red Lake greenstone belt. A scenario, to be advocated in the section entitled Tectonic Interpretation and Outstanding Problems, involves an allochthonous or parautochthonous, northward transport of the Confederation and younger assemblages (St. Joseph and English River), with limited, parautochthonous displacement of older assemblages toward the north.

Lake St. Joseph Greenstone Belt

The Lake St. Joseph greenstone belt records a critical part of the structural history in that the events postdating 2713 Ma volcanism include thrust stacking of volcanic assemblages, crustal exhumation and erosion followed by orogen-parallel folding, plutonism and later transcurrent faulting. The basic features of this sequence of events are given here and summarized in Figure 6.18b.

The D_2 structures of the southern Birch–Uchi greenstone belt may correspond with the orogen-parallel folding and transcurrent shear fabric of the Lake St. Joseph greenstone belt where Clifford (1969, 1972) documented several phases of deformation; the authors have coupled his observations with the observations of Berger (1981) and Stott et al. (1987a, 1987b) plus age determinations by the second author, F. Corfu, to postulate here a sequence of events as summarized below and in Figure 6.18b. The initial assembly of the Lake St. Joseph greenstone belt involved tectonic repetition (*see* Figure 6.6) of 2 assemblages: the Confederation assemblage, comprising high-magnesium tholeiitic basalt flows and overlying 2734 million-year-old calc-alkalic rhyolite, and the St. Joseph assemblage comprising at least 2 cycles of volcanism, one of tholeiitic basalt capped by a 2723 million-year-old dacite to rhyolite pyroclastic edifice and the second of 2713 million-year-old basalt-andesite-dacite. In the western part of Lake St. Joseph, these 2 assemblages, separated by a hiatus, formed the core of a stratovolcano (Clifford and McNutt 1971). The evidence from age determinations (Corfu and Stott 1989) of volcanic cycles in this area and trace plus rare earth element geochemistry (Clifford and McNutt 1971; Stott, unpublished data) show that the 2 assemblages were repeated by fault imbrication so that in Figure 6.6, Cycle 1 (Confederation assemblage), which shares the same volcanic vent system as the younger and overlying Cycle 2 (St. Joseph assemblage), recurs as Cycle 3, overlain by Cycle 4, the St. Joseph assemblage. The volcanic units in all cycles show dominant way-up directions towards the north and east. This tectonic stacking of strata was rapidly followed by erosive exhumation of continental crust exposed from the north, debris from which forms the unconformably overlying sedimentary Eagle Island assemblage. This assemblage includes deposits of both clastic detritus and thick iron formation beds (a submarine channel fan?) adjacent to turbidite of the English River assemblage (fore-arc basin?). Deposition of the Eagle Island assemblage may have been accompanied by soft-sediment slumping, subsidence, dewatering and folding of the proximal basin. This period of sedimentation was followed by collision between the Uchi–Sachigo and the Winnipeg–Wabigoon superterranes across a suture zone (the tectonic expression of a collision zone; Howell 1989) underlying what may have been a chain of English River basins to form an accretionary prism. The collision stage is characterized by the presence of orogen-parallel folds, including the anticline that folds the Eagle Island and older assemblages; the abundance of elongate, orogen-parallel granitic plutons near the subprovince boundary in the Red Lake greenstone belt (*see* Figure 6.2; *see also* Map 2542, map case); and the prevalence of eastward-striking schistosity and bedding in the southern half of the subprovince. This postulated sequence of events culminated in the emplacement of late to posttectonic granodiorite plutons, including the granodioritic Blackstone pluton at 2702±1 Ma, which was cut by the Sydney Lake–Lake St. Joseph Fault and shows kinematic evidence of limited dextral, transcurrent displacement (*see* Figures 6.22a to 6.22d).

Meen–Dempster Greenstone Belt

The Meen–Dempster greenstone belt illustrates the use of stretching lineations in regional-scale mapping to establish the extent of structural domains, including contact strain aureoles produced by the emplacement of plutons. A sequence of structural events is outlined here.

The Meen–Dempster greenstone belt, composed of 2 volcanic assemblages that face dominantly southward, has been subdivided by Stott and LaRocque (1983a, 1983b) and Stott and Wilson (1984) into 2 broad structural domains,

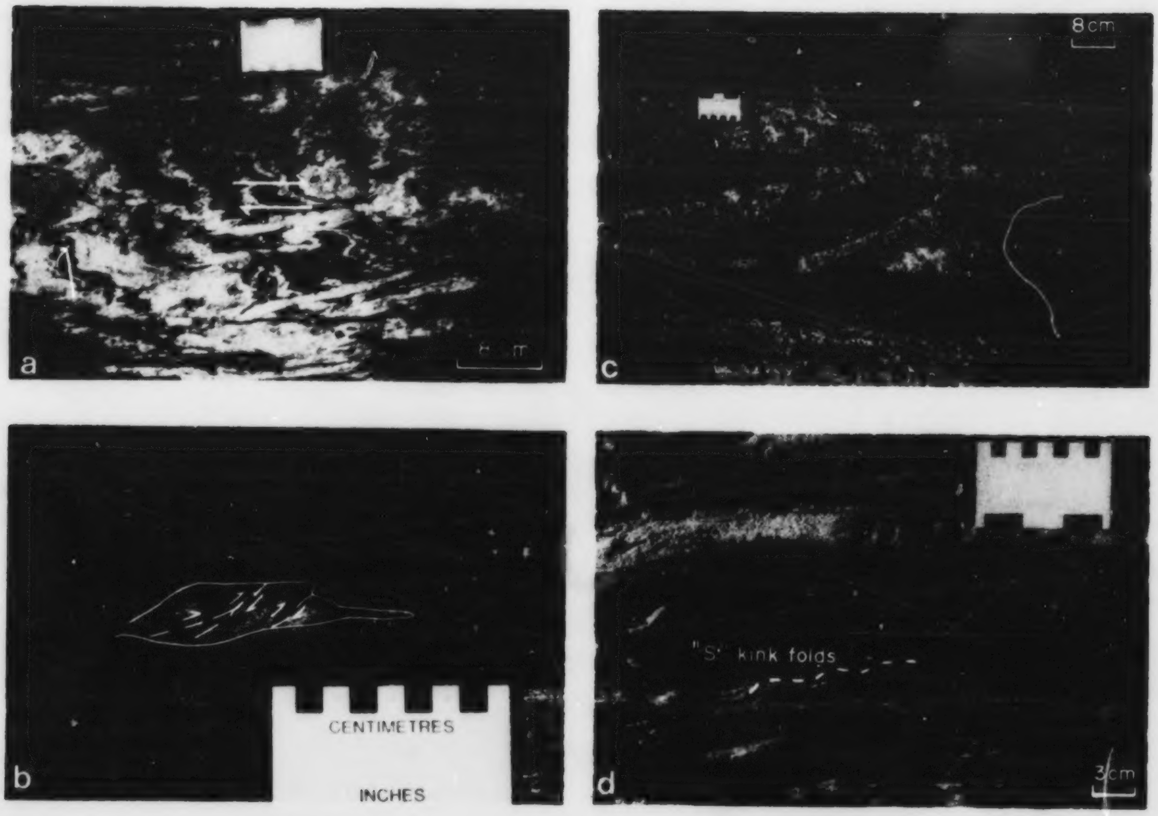


Figure 6.22. Kinematic indicators showing right-handed transcurrent sense of movement in the Sydney Lake–Lake St. Joseph Fault: **a)** Z-folded layers of banded-magnetite iron formation and wacke-sandstone in the fault zone on western Lake St. Joseph; **b)** transcurrent duplex stacking of a vertical wacke-sandstone layer in banded iron formation-wacke sequence on western Lake St. Joseph. Transcurrent thrusting occurred along the iron formation-wacke interface and indicates that displacement was accommodated locally by thickening of layered sequences; **c)** duplex stacking by transcurrent thrusting in English River assemblage-wacke sandstone and siltstone units near Slate Lake, south of the Birch–Uchi greenstone belt; **d)** s-kink folds in schistose, magnetite-iron formation–wacke-sandstone sequence on western Lake St. Joseph. Kink folds appear prominently in and near the Sydney Lake–Lake St. Joseph Fault.

which are locally overprinted by younger, contact strain aureoles bordering late granitic plutons. The 2 domains (*see* Figure 6.11) both display steeply northward- dipping schistosity that closely parallels the overturned volcanic bedding. This belt is, however, locally folded close to external batholiths (Stott and LaRocque 1983; Rodd and Hutchinson 1991), evident from reversals in stratigraphic younging directions near the batholiths. There is otherwise little observable evidence of folding. There is a moderate state of strain across most of the belt with intense, apparently irrotational flattening strain in some narrow deformation zones and rotational shear strain in other zones that could be related to the Bear Head fault zone along the northern margin of the belt.

A common schistosity across these 2 domains inhibits their distinction, but one can recognize these domains (*see* Figure 6.9) from the orientations of stretching lineations (e.g., of varioles, clasts and mineral aggregates) that differ in each domain. These are accompanied by parallel mineral lineations that are more widely distributed and readily observed. Suitable minerals include prismatic amphibole crystals, quartz and feldspar grains and the streaks of phyllosilicates.

The oldest recognizable strain fabric occupies the D_1 domain and is characterized by westward-plunging lineations on the plane of schistosity in rocks to the east and south of the Graniteboss stock. The western half of the belt is marked by eastward-plunging lineations, with a typically 30° to 40° plunge. This fabric can also be traced into the southern part of the Dobie Lake pluton, which typifies the D_2 domain, and can be traced into a 1.5 km wide shear zone that defines part of the domain's southern boundary. The size and location of the D_2 domain is notably coincident with the size and presence of the Dobie Lake pluton north of the belt. In addition, this domain is bounded by 2 high-strain deformation zones and appears from paleomagnetic evidence (Hale and Lloyd 1990) to have rotated, as a block, clockwise. These considerations coupled with the consistent eastward plunge of mineral lineations in the slightly metamorphosed Dobie Lake pluton lead one to postulate the following: the D_2 domain is a broad zone of strain produced by the Dobie Lake pluton as the pluton was translated and rotated south and southeastward during *circa* 2700 million-year-old regional orogenic shortening; the pluton deflected that part of the belt southward and induced a clockwise block rotation. The overall mechanics can be likened to squeezing a pebble in a muddy matrix with the enveloping matrix (D_2-part of the greenstone belt) absorbing more than its share of the bulk strain against a comparatively rigid pluton. The orogenic shortening produced a tectonic schistosity in the 2722 million-year-old Kawashe Lake stock (Corfu and Stott 1989) and the Billett assemblage, but predated the emplacement of the posttectonic 2701 million-year-old Graniteboss stock (F. Corfu, unpublished data).

The subsequent strain effects of younger plutons upon the greenstone belt vary; for example, the geometric pattern of flattened strata in the greenstone envelope around the Obaskaka pluton implies that the pluton may have expanded during emplacement. The Billett assemblage, for example, is thickened where it strikes towards the east side of the pluton and becomes attenuated as it bifurcates around it to the north and south (Sage and Breaks 1982). The tabular and crescent-shaped intrusions (e.g., the Osnaburgh pluton) have narrow contact strain and metamorphic aureoles; one can recognize the strain aureoles by the change in orientation of stretching directions in the greenstone envelope, from a regional-scale trend and plunge to a more local orientation, adjacent to the pluton, which conforms to the stretch lineations produced within the pluton during its emplacement. One example of the strain effects can be observed at the west end of the Meen–Dempster greenstone belt where detailed lineation measurements show a progression in lineation orientations as one approaches the late to posttectonic external pluton (labelled Southern pluton in Figure 6.20). The strain envelope is imprinted upon the existing regional schistosity in the belt. With few exceptions, only a single lineation is preserved on or parallel to a schistosity surface so that we see a progression of orientations of prismatic amphiboles, micaceous streaks, mineral aggregates, vesicles and clasts.

A right-handed transcurrent shear zone, with a shallow southeastward-plunging direction of shear, follows the length of the belt along its northern margin. It is well expressed within the margin of the Dobie Lake pluton, varying from parallel sets of narrow mylonite zones to a broader zone 1.5 km wide with a range of high to moderate shear strain. This zone is part of the Bear Head fault zone, which is described further in the section entitled Shear Zones and Faults. It probably has a long episodic history, initially as a zone conveying ductile shear during the late stage of cratonization of the northern Superior Province (the "Uchian phase" of the Kenoran Orogeny; *see* Sequence of Events in the Uchi Subprovince), and more speculatively, later as one of a number of reactivated zones of brittle relief in the interior of the Archean craton during creation of the 1800 million-year-old Trans–Hudson Orogen, which adjoins the Superior Province to the northwest.

The timing constraints and the overall sequence of structural and magmatic events in this belt are summarized in Figure 6.18c.

Pickle Lake Greenstone Belt

The Pickle Lake greenstone belt has been subdivided into 3 structural domains (*see* Figure 6.10a) by Stott et al. (1989a). The oldest deformation fabric occurs in Domain 1 and comprises a steeply northwestward-dipping schistosity that typifies the generally northward-dipping schistosity prominent in most of the major belts of the Uchi Subprovince. The domain is characterized by a steep, northeasterly plunging lineation of amphibole and other mineral grains and shaped ellipsoids of varioles, clasts and mineral aggregates. Domain 1 includes most of the mineralized deformation zones of the Pickle Crow gold deposits. The gold-quartz veins in the Domain 1 deformation zones are steeply folded. Domain 2 comprises a grouping of spatially restricted contact strain aureoles surrounding, or adjacent to, late felsic plutons such as the Pickle Lake stock and the Ochig Lake pluton. Features of pluton-induced strain aureoles are summarized later. Domain 3 is characterized by a schistosity that dips moderately to the northwest, away from the adjacent 2713 million-year-old Second Loon pluton in the Seach–Achapi batholith. The westward-plunging lineations of this domain are parallel to the lineations of the Second Loon pluton. One may infer that the fabric of Domain 3 represents a wide contact strain aureole related to the emplacement of the Second Loon pluton. The domainal structural subdivision of the Pickle Lake greenstone belt reflects an attempt to distinguish areas containing older structures from parts of the belt that are dominated by the deformational effects of external and internal pluton emplacements. Radiometric age determinations of zircons in some of the plutons allow constraints on the timing of structural events in this and other belts as summarized in Figures 6.18a to 6.18d.

REGIONAL NON-PENETRATIVE STRUCTURES

Shear Zones and Faults

Mapping of the belts on different scales over the years has revealed numerous deformation zones (broad zones of rock containing discrete faults and shear zones), which were active at different stages in the evolution of the subprovince. In some cases, a part of the kinematic history has been determined, as exemplified in the Red Lake greenstone belt where deformation zones (Durocher and Burchell 1983; Durocher et al. 1987, 1991) have been the focus of gold mineralization studies and consequently placed in a historical sequence of events (Andrews et al. 1986; Corfu and Andrews 1987; Table 6.10). Other examples were described from several greenstone belts in the previous subsection.

The time-constraining evidence for gold mineralization in the Uchi Subprovince is based on observations by Andrews et al. (1986) and Stott et al. (1989a, 1989b), plus published and unpublished age determinations by F. Corfu (Royal Ontario Museum). There is a significant difference in the timing of gold mineralization in the Red Lake greenstone belt and the Pickle Lake greenstone belt, inferred from the age of deformation events and crosscutting relationships. The evidence supports the conclusion that there were different episodes of gold mineralization in this subprovince, arising from different regional orogenic events (Kenoran and earlier) and from pluton emplacements. Additional discussion is given in Fyon, Breaks et al. (this volume).

Investigations show that the deformation zones in the Uchi Subprovince have different origins. Indeed, on a broader scale, one could informally assign faults and shear zones in northwestern Ontario to 5 categories (Stott 1991), based on geochronology, structural chronology, and spatial relationships to greenstone belts, plutons and subprovince boundaries. These categories are:

1 intrabelt terrane-boundary faults
2. intrabelt faults and deformation zones
3. broad domains of deformation containing numerous shear zones
4. transcrustal megafaults
5. pluton-induced shear zones

These are interpretive categories and examples of these deformation zones in the Uchi Subprovince are discussed below.

TYPES 1 AND 2. INTRABELT TERRANE-BOUNDARY FAULTS AND OTHER INTRABELT DEFORMATION ZONES

Of particular note, because of their significance to gold mineralization, are the faults and shear zones spatially associated with hydrothermal alteration zones within greenstone belts, such as the Red Lake greenstone belt (*see* Figure 6.4) described by Pirie (1982); broad domains of deformation as seen in the Meen–Dempster greenstone belt (*see* Summary for a description); and trans-Superior megafaults such as the Sydney Lake–Lake St. Joseph Fault on the Uchi–English River subprovince boundary.

Gold prospects and mines (former and active) are especially abundant in broad bands in the Red Lake greenstone belt. Several workers (e.g., Andrews et al. 1986; Hugon and Schwerdtner 1988) have shown these broad bands to coincide with deformation zones (*see* Figure 6.4) that contain a multiplicity of narrow high-strain ductile and brittle-ductile shear zones (*see* Figure 6.17a). The structural relationships in these zones are complex in detail and are well displayed in the gold mines of this belt. The origin of these shear zones in the Red Lake greenstone belt is a controversial subject. Some workers, such as Andrews et al. (1986), attributed the initiation of the zones to the emplacement of the adjacent external plutons and viewed the age-constraining evidence as supporting a sequence (shown in Figure 6.17b) of magmatism, deformation and gold mineralization. However, some of the deformation zones coincide with boundaries between tectonic assemblages and accordingly, part of their displacement history may correspond to a regional orogenic event on a broader scale that exceeds the deforming role of the individual plutons. For example, the boundary between the Balmer and Confederation assemblages in the Red Lake greenstone belt strikes northeastward from the Killala–Baird batholith (*see* Figure 6.4) and coincides with the Flat Lake–Howey Bay deformation zone. The assemblage boundary is cut by and thus predates the 2718 million-year-old Dome stock. The deformation zone following this boundary is overprinted by a thermal aureole produced by the Killala–Baird batholith (Andrews et al. 1986), which limits the minimum timing of reactivation of this boundary to 2704 Ma, the age of the batholith phase close to the belt. The reactivation of this deformation zone and other faults along assemblage boundaries across the belt may have coincided with the emplacement of the external plutons, as Andrews et al. (1986) have suggested, to form other intrabelt faults and shear zones of type 2.

TYPE 3. BROAD DOMAINS OF DEFORMATION CONTAINING NUMEROUS SHEAR ZONES

Type 3 domains are typified by the northern half of the Shebandowan greenstone belt in the Wawa Subprovince (*see* Williams, Stott, Heather et al., this volume), but can also include strain domains that were indirectly induced by plutons during regional orogenic shortening; an example from the Meen–Dempster greenstone belt is described later and illustrated in Figure 6.9.

TYPE 4. TRANSCRUSTAL MEGAFAULTS

The fourth type of fault comprises the major trans-Superior fault zones, best exhibited in the northern half of the Superior Province. The prominence of eastward- and southeastward-striking faults with consistent right-handed sense of displacement has been taken by most workers (e.g., Schwerdtner et al. 1979; Park 1981) to reflect a late phase of regional south-southeast-directed shortening. As the deformation during the Kenoran Orogeny became progressively more advanced, partitioning into faults and shear zones was accompanied by block rotations (e.g., Hale and Lloyd 1990). Most of the faults traversing the Uchi Subprovince strike southeastward. The following is a description of 2 examples of these faults examined by Osmani and Stott (1988) plus the east-striking Sydney Lake–Lake St. Joseph Fault along the Uchi–English River subprovince boundary.

The Bear Head fault zone follows the boundary between the Berens River and Sachigo subprovinces of Card and Ciesielski (1986) and can be traced from Lake Winnipeg in Manitoba; its main branch continues into the North Spirit Lake greenstone belt (Stone 1990), but it is also interpreted to continue, through a horsetail-like splay, southeastward towards the Meen–Dempster greenstone belt. In the Uchi Subprovince, this fault crosses plutons of hornblende-bearing potassium-feldspar megacrystic granite; it appears on the aeromagnetic maps as a narrow line of high magnetic intensity crossing the plutons, and owes its expression to shear-induced alteration of hornblende to magnetite, chlorite and epidote. The rocks within the fault zone have been deformed to protomylonites, which enclose very fine-grained mylonite bands. Mineral lineations and the long axes of mineral aggregates typically plunge 5° to 30° to the southeast. Various kinematic criteria—progressive clockwise rotation of the schistosity plane towards the shear plane, ductile shear bands, and asymmetric pressure shadows of porphyroclastic potassic feldspar megacrysts—demonstrate a right-handed sense of shear on the horizontal plane (Figures 6.23a and 6.23b)).

The Bear Head fault zone can be traced as a zone, typically 1 km wide along the northern margin of the Meen–Dempster greenstone belt, increasing to 3 km wide in the vicinity of MacDowell Lake in the Berens River Subprovince (Osmani and Stott 1988). It is typical of numerous curviplanar faults in northwestern Ontario, following close to greenstone belt margins. Since these faults transect 2.7 billion-year-old plutons, they represent late displacements during and subsequent to the main period of Kenoran orogenesis.

The North Caribou Lake–Totogan Lake shear zone displays a right-handed transcurrent shear sense as demonstrated by the pronounced deflection of regional magnetic trajectory lines into the southeastward direction of shear and by kinematic details of the rock fabric (Osmani and Stott 1988). It closely follows a chain of greenstone belts lying in both the Sachigo and Uchi subprovinces. It has, therefore, been reinterpreted by Thurston, Osmani et al. (this volume) as a subprovince or terrane-bounding structure. These southeast-striking faults may be attended by parallel shear zones that follow the bedding and/or schistosity within the adjacent greenstone belts.

From airphoto interpretation, Parkinson (1962) postulated numerous fault zones in the Superior Province, including the Lake St. Joseph fault. This fault has now been shown by reconnaissance mapping (e.g., Williams 1988) to be continuous with the Sydney Lake fault south of Red Lake. The features of this fault have been documented by Stone (Stone 1977, 1981), Breaks et al. (1978) and Schwerdtner et al. (1979). The east-striking Sydney Lake–Lake St. Joseph fault zone is steeply dipping and can be traced along the Uchi–English River subprovince boundary for at least 440 km from Lake Winnipeg in Manitoba to Pashkokogan Lake where the fault joins 2 other fault zones (the Miniss River and Pashkokogan Lake–Kenoji Lake faults) that cross the English River Subprovince. The Sydney Lake–Lake St. Joseph fault zone, typically 1 to 2 km wide, is generally a continuous structure (e.g., south of Lake St. Joseph), but in places (e.g., south of the Red Lake greenstone belt) it contains a braided network of narrow shear zones. Kinematic indicators and evidence of transcurrent displacement across this fault show a consistent right-handed sense of movement. If the direction of regional compression was constant at the time of transcurrent faulting, then dextral transcurrent movement across northeast-striking splays of this fault require that the maximum compressive stress must have been along a northwest-trending (~300°) axis (Stone 1977).

Kinematic indicators of right-handed displacement include Z-folds, shear bands and sygmoidal asymmetry in the compositionally layered rocks (*see* Figure 6.22). Some asymmetric folds are 500 m wide across the limbs (e.g., at the southeastern end of the Bee Lake belt). Stone (1977) estimated 6 km of dextral displacement across part of the fault south of the Bee Lake belt, in accord with estimates made at Lake St. Joseph by Stott et al. (1987a, 1987b). There is a progressive increase in strain toward the centre of the fault zone. These intensely schistose rocks are overprinted by scattered concentrations of brittle kink bands (e.g., Clifford 1968), products of localized east-west shortening, which accompanied late stage transcurrent displacements within this fault zone. Brittle, localized, post–Archean reactivation of these fault zones likely occurred across Ontario, based on Proterozoic ages for fracture-zone materials (Kamineni et al. 1990) from the Lac du Bonnet batholith at the Bird River–Winnipeg River subprovince boundary in Manitoba; such brittle structures, as well as the dike swarms, are part of the post–Archean disturbances of the Superior Province as the North American continent was growing.

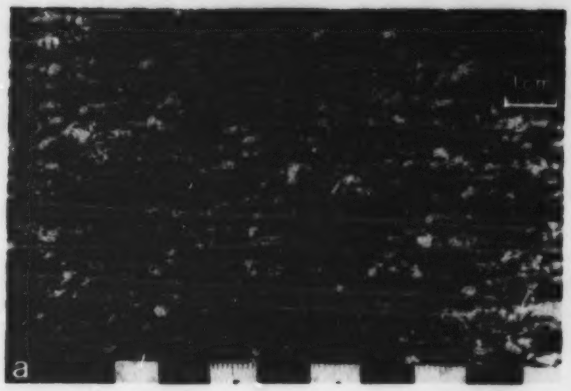


Figure 6.23a. Potassium feldspar-megacrystic monzogranite subjected to right-handed simple shear in the Bear Head fault zone. Right-handed asymmetry of local schistosity rotated into shear planes and rolled potassium feldspar augen highlight the sense of movement. Photo taken at Kamungishkamo Lake, northeast of the Birch–Uchi greenstone belt. Centimetre scale at bottom.

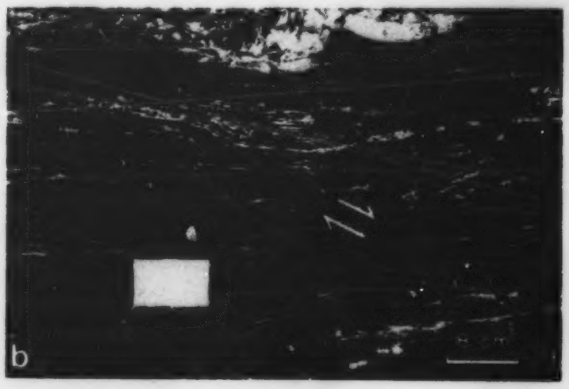


Figure 6.23b. Shear-banded chloritic schistose basalt, showing an asymmetry consistent with right-handed transcurrent displacement in the Bear Head fault zone, along the northern margin of the Meen–Dempster greenstone belt.

TYPE 5. PLUTON-INDUCED SHEAR ZONES

The active intrusion of plutons can produce contact strain aureoles that include discrete shear zones. These are described in the following section.

PLUTON-INDUCED STRUCTURES

Penetratively strained rocks in a greenstone belt may arise from regional stresses applied to the belt during its assembly or during subsequent orogenic shortening; or from the forceful emplacement of a pluton. This section looks at the role of the plutons in producing some of the structures and orientations of strata observed in the greenstone belts. This role can be assessed in terms of strains actively imposed upon the host rocks surrounding a pluton (i.e., contact strain aureoles); deflections of strata by the emplacement of a pluton, but without significant penetrative strain imposed by the pluton; and indirectly induced structures arising from the apparent use of the pluton as a passive tool by orogenic shortening stresses to compress the strata in an adjacent greenstone belt, conceptually analogous to a pebble compressing an adjacent mudstone layer. Two of these 3 roles are described briefly, with emphasis placed on contact strain aureoles because of their potential significance for some types of gold deposits. An example of "passive" deflections of strata is not evident in the Uchi Subprovince.

Contact Strain Aureoles

Some plutons have forcefully intruded into place and during their emplacement, they imposed a penetrative strain upon the host rocks. This zone of deformation in the rocks that surround the pluton is called a contact strain aureole. The width of the aureole is related to the nature of the pluton emplacement as well as the size of the pluton since plutons may to some degree stope their way into place and to that degree, they may impose relatively less strain upon the wall rocks. A spectrum of emplacement mechanisms appears to be represented among the plutons of the subprovince since some of the larger, late tectonic bodies stoped their way into place and left no contact strain aureole upon their surrounding host rocks. The Williams Lake batholith, north of central Uchi Subprovince, is a good example.

However, most of the late plutons within and adjacent to the greenstone belts display a contact strain aureole and usually a narrower contact thermal metamorphic aureole. The aureole is typically revealed as a change from regional greenschist metamorphism in the basaltic envelope to amphibolite grade near the pluton. The creation of the strain aureole in the host rock that borders the pluton can be attributed to the directed stress imposed by the active emplacement, and spreading or ballooning of a pluton when it can rise no further in the crust. The extent of these aureoles in the surrounding or adjacent greenstone belt can be gauged where the orientation of stretching lineations, resulting from regional, tectonically induced attitudes, changes to attitudes that correspond closely to the last increment of strain frozen in the pluton's fabric as the pluton was emplaced. Care is needed to ensure that the pluton fabric is indeed indigenous to the emplacement of the pluton (e.g., Paterson et al. 1989).

A generic strain aureole is shown in Figure 6.24 as it relates to some discrete zones of gold mineralization described in the section entitled Deposit Types. Some examples of strain aureoles produced by plutons in the Uchi Subprovince are summarized in Table 6.8; the fabric elements arising from strain effects imposed by the Ochig Lake pluton in the Pickle Lake greenstone belt are illustrated in Figures 6.10 and 6.15. Strain aureole fabrics can be traced for several kilometres into the surrounding greenstone belts and are accompanied by a thermal aureole of amphibolite grade.

The contact strain envelope of the Ochig Lake pluton postdates other strain domains of the Pickle Lake greenstone belt and is the main evidence to indicate that regional shortening during the 2.7 million-year-old Kenoran Orogeny did not produce a penetrative strain fabric in the Pickle Lake greenstone belt. The penetrative deformation, seen in the southern half of the Uchi Subprovince and associated with the Kenoran Orogeny, may not have affected most of the region in the northernmost part of the Uchi Subprovince, including the Pickle Lake greenstone belt and northern Minimiska–Fort Hope greenstone belt. The Pickle Lake greenstone belt may therefore lie north of an "orogenic front" of penetrative Kenoran deformation. This observation is consistent with the absence of penetrative Kenoran orogenic deformation in much of the North Caribou terrane farther north (*see* Thurston, Osmani et al., this volume).

Indirectly Induced Structures

Rotation of supracrustal strata may arise from the physical rotation of the pluton itself during later orogenic shortening. A broad deformation zone that appears related to an individual pluton, such as the D_2 zone south of the Dobie Lake pluton described earlier, may have an alternative origin than the directed stress caused by the active emplacement of the pluton. The Dobie Lake pluton may have indirectly produced the D_2 domain of the Meen–Dempster greenstone belt as the pluton was translated and rotated during regional orogenic shortening. The resulting deformation would have been partitioned into broad zones of flattening and extension by irrotational pure shear and narrow, intervening high-strain zones characterized by rotational shear. Such an indirectly induced structural process is envisaged to be no different in kind than those arising from the translation and rotation of crustal blocks, although the effects of crustal block displacement are more commonly known to be focussed into ductile shear zones.

SUBPROVINCE BOUNDARIES

The southern boundary of the Uchi Subprovince is a tectonic boundary, a fault contact against the sedimentary English River assemblage. In contrast, a batholithic belt—the Berens River Subprovince—lies to the north of the Uchi Subprovince. The northern boundary of the Uchi Subprovince is an artifact of a previous tectonic subdivision by Card and Ciesielski (1986) and earlier workers. The Uchi Subprovince merges into the Berens River batholitic belt without a fault boundary to separate the two, as noted earlier. One can trace the structural grain of tonalitic gneisses, striking north and northeastward from Uchi Subprovince across the plutonic belt. The distinction of the Berens River Subprovince remains only a lithological one.

This batholitic belt has an undetermined history of growth as it became exhumed during the Kenoran Orogeny;

by its overprinting relation masking the early connecting units between the Uchi and Sachigo subprovinces, by the presence of multiple suites of intrusive units and by its size. If considered to extend discontinuously from the Berens River Subprovince to the Bienville Subprovince in Quebec (Card and Ciesielski 1986), it superficially resembles a more deeply exhumed example correspondent to the Coastal Batholith of the Western Cordillera of Peru (Pitcher et al. 1985), although there is no evidence of a progressive change in plutonic composition across the batholitic belt north of the Uchi Subprovince. Descriptions of the Berens River batholitic belt are given in Thurston, Osmani et al. (this volume).

Most of the southern boundary separating the Uchi and English River subprovinces is marked by a sharp change in rock types and metamorphic grade across the Sydney Lake–Lake St. Joseph Fault, observed from Manitoba (McRitchie and Weber 1971a) into Ontario (Breaks et al. 1978) to at least the vicinity of Lake St. Joseph (Clifford 1969; Stott et al. 1987a, 1987b). However, south of the Miminiska–Fort Hope greenstone belt, the boundary is less clearly defined and, like the southernmost part of the Birch–Uchi greenstone belt, it resembles the pattern of interleaved panels of mafic volcanic and clastic sedimentary units of the Beardmore–Geraldton Belt that characterize the accretionary-prism hypothesis postulated by Williams (1990) for the southern margin of the Wabigoon Subprovince. The inference, yet to be tested, is that parts of the Uchi–English River subprovince boundary may preserve comparable evidence as a transitional zone into an accretionary prism. In the western part of Lake St. Joseph, there is a well-preserved assemblage of clastic and iron formation sedimentary rocks

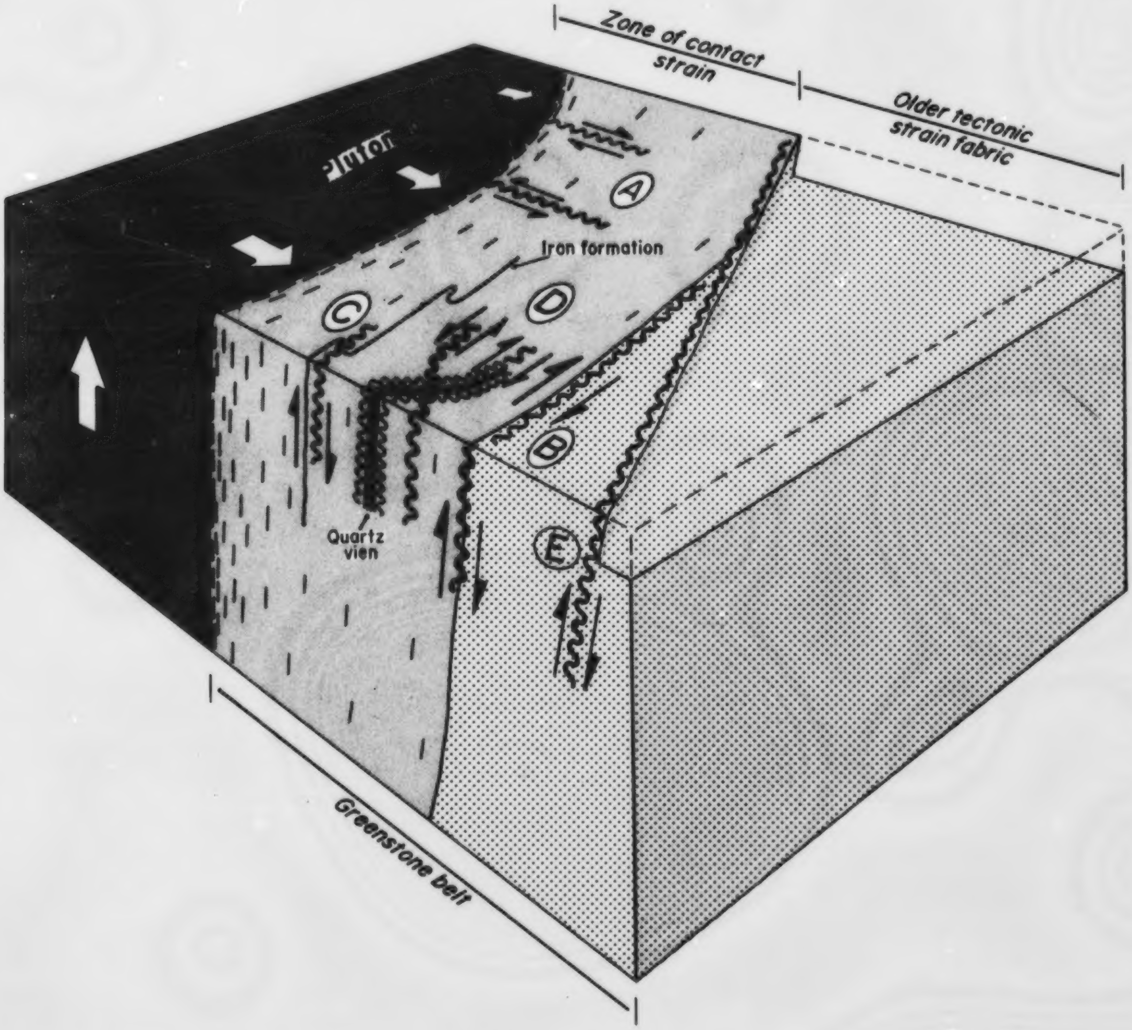


Figure 6.24. Schematic block diagram of a contact strain aureole around a pluton. For an explanation of types A to E shear zones, *see* Gold Deposits. Two types of gold-bearing shear zones (types C and D) may be represented in the Pickle Lake greenstone belt.

Table 6.8. Examples of plutons that induced contact strain aureoles.

Pluton	Greenstone Belt Affected	Width of Strain Aureole	Fabric Characteristics	References
Walsh Lake (2699 ±1 Ma)	Red Lake	7 km	- northward-dipping schistosity from antiformal crescentic pluton - mineral lineations and long axes of pillows plunge downdip - outermost perimeter of aureole marked by shear zone.	Andrews et al. (1986) Noble (1989) Hugon and Schwerdtner (1988)
Blackstone (2701 ±1 Ma)	West Lake St. Joseph	1.5 km	- flattened pillows - north-plunging lineations shift to southeast-plunging outside aureole	Clifford (1969) Stott et al. (1987a)
Ochig Lake	Pickle Lake	up to 3 km	- outward-dipping schistosity from domical pluton and corresponding schistosity in greenstone envelope - mineral lineation downdip in pluton and greenstone envelope (*see* Figure 6.15)	Stott et al. (1989a)

documented by early mapping geologists and described by Clifford (1969), Meyn and Palonen (1980) and Berger (1981).

The Sydney Lake–Lake St. Joseph Fault forms an interface across which the metamorphic grade changes sharply, yet a significant part of felsic volcanic stratigraphy southeast of Lake St. Joseph (*see* Figure 6.3) has been transected and removed at the Sydney Lake–Lake St. Joseph fault boundary. Since there has been no lithological correlation indicating lateral transfer of this missing volcanic strata, the fault was probably initiated as a north-verging thrust. This thrust locally removed part of the southern margin of the Uchi Subprovince and carried a deeper section of the English River rocks upward in juxtaposition to the greenschist grade rocks of western Lake St. Joseph, for example. This fault subsequently served as a dextral transcurrent shear zone in the late stage of Archean crustal history (post–2702 Ma) in this region.

SUMMARY

Although the northern part of the Uchi Subprovince contains a record of deformation that preceded the main phase of the Kenoran Orogeny, the subprovince is dominated by a sequence of events related to regional orogenic shortening and telescoping of the crust during what is best viewed as a collision between the Uchi–Sachigo superterrane and the Winnipeg River–Wabigoon superterrane to the south; the orogenic deformation is accompanied, and locally masked, by the strain effects of pluton emplacement. The sequence of events includes thrust stacking of steeply dipping strata, juxtapositioning of some assemblages, folding and accompanying shearing along the assemblage boundaries, and later transcurrent fault displacement as the superterranes converged at an oblique angle along a north-northwest axis. Features of the structural sequence are summarized in the section entitled Sequence of Events in the Uchi Subprovince.

Metamorphism

METAMORPHIC FACIES AND CONDITIONS

The regional pattern of metamorphism in the Uchi Subprovince has been reviewed by Thurston and Breaks (1978) and is shown in Figure 6.25. Only a very general appreciation of the metamorphic record is documented in this subprovince. Owing to the volume of late tectonic granitic plutons, a complex history of metamorphic events exists which arises from episodes of burial, crustal thickening, thrusting, baking by felsic intrusions of all sizes and ages, and late hydrothermal alterations focussed along shear zones. Overall, the greenstone belts have been subjected to low pressure Abukuma-type metamorphism ranging from very low to high temperature grades. The belts are principally characterized by low grade (greenschist facies) metamorphic mineral assemblages but contain amphibolite facies mineral assemblages in areas close to the major felsic plutons, both internal and external to the belts, and in some areas close to the English River subprovince boundary. Evidence for very low grade

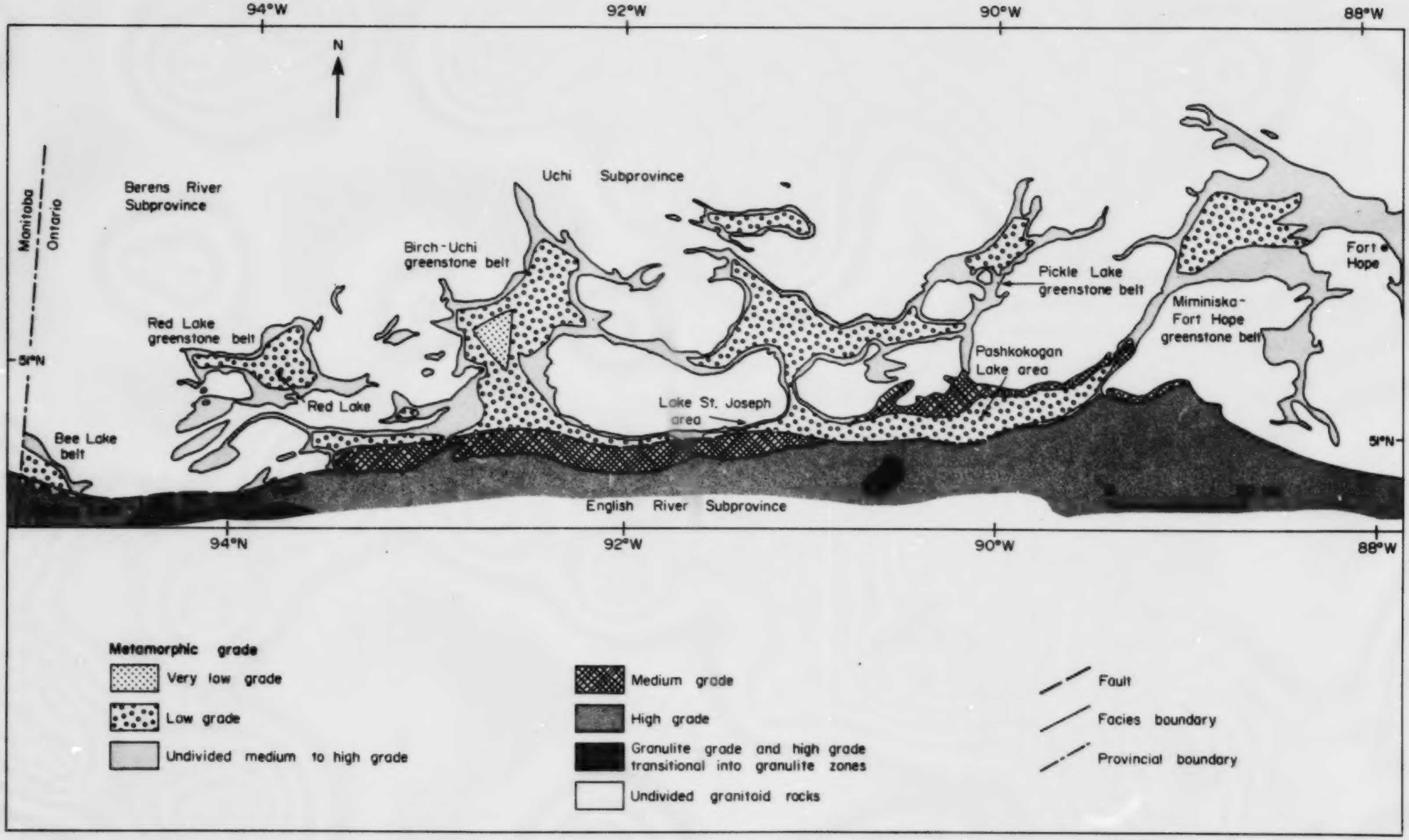

94°W
92°W
90°W
88°W
N
Manitoba
Ontario
51°N
Berens River Subprovince
Uchi Subprovince
Birch-Uchi greenstone belt
Red Lake greenstone belt
Red Lake
Bee Lake belt
Lake St. Joseph area
Pickle Lake greenstone belt
Pashkokogan Lake area
Miminiska-Fort Hope greenstone belt
Fort Hope
English River Subprovince
94°N
92°W
90°W
88°W
Metamorphic grade
Very low grade
Low grade
Undivided medium to high grade
Medium grade
High grade
Granulite grade and high grade transitional into granulite zones
Undivided granitoid rocks
Fault
Facies boundary
Provincial boundary

Figure 6.25. A generalized map of the distribution of metamorphic zones in the greenstone belts of the Uchi Subprovince and northern English River Subprovince (*modified from* Thurston and Breaks 1978).

Table 6.9a. Metamorphic mineral assemblages in rocks of the Earngey–Costello area, Birch–Uchi greenstone belt (*modified from* Thurston 1985b).

	Metasedimentary Rocks	**Felsic Metavolcanic Rocks**
Greenschist Facies	All rocks contain quartz + albite + biotite, with: • muscovite + andalusite ±epidote • epidote + muscovite + chlorite • epidote • muscovite + chlorite + stilpnomelane ±epidote • muscovite + staurolite (relict)	All rocks contain albite + quartz + microcline, with: • biotite + epidote + carbonate • muscovite + biotite + chlorite + epidote + carbonate • biotite + chlorite + epidote ±carbonate • biotite + tremolite + epidote ±chlorite • chlorite + muscovite + epidote • muscovite + biotite + epidote + carbonate • muscovite + chlorite + stilpnomelane + epidote + carbonate • tremolite + chlorite + epidote
Amphibolite Facies	All rocks contain andesine-oligoclase + quartz + biotite, with: • chlorite ±cordierite • muscovite + chlorite + garnet + cordierite ±epidote • cordierite + muscovite • muscovite + cordierite + garnet + sillimanite • microcline + cordierite + muscovite • muscovite + cordierite + andalusite ± garnet	• plagioclase + quartz + microcline + actinolite
Gabbro and Diorite		
Peridotite and Dunite		

metamorphism is restricted to the central part of the Birch–Uchi greenstone belt (Thurston 1985b, p.62), where relict pyroxene in basaltic flows and relict snowflake texture in rhyolitic rocks are observed. Otherwise, this belt contains metamorphic mineral assemblages (Table 6.9a) that typify greenschist to amphibolite facies conditions described in other greenstone belts of this subprovince, such as the: Bee Lake Belt (Shklanka 1967); northern Birch–Uchi greenstone belt (Beakhouse 1989; Good 1988); Miminiska–Fort Hope greenstone belt (Wallace 1981a). Local thermal gradients are not always readily attributed to plutons, although the thermal source may lie at depth. For example, a pattern of low pressure metamorphism, with a gradient of decreasing temperature (Table 6.9b) in the sedimentary assemblage on Miminiska Lake, has been shown by Wallace (1981a, modified after Sharpe 1979) to be symmetrically centred on the assemblage without a clearly attributable thermal source.

Contact thermal metamorphic aureoles, most prominently consisting of hornblende + plagioclase in basaltic flows, are typically several hundred metres wide, but vary depending on the size of the adjacent pluton, its geometry at depth, available fluids and its cooling rate. The thermal effect of the external plutons is most apparent in the smaller greenstone belts, where broad zones of amphibolite facies metamorphism occur and only a limited central part of the belts may contain greenschist facies rocks as in the Bee Lake belt. Significant breadth of contact metamorphism was documented by Andrews et al. (1986) for the eastern part of the Red Lake greenstone belt where amphibolite facies metamorphism extends up to 9 km from the Trout Lake batholith to the vicinity of the Arthur W. White gold mine (*see* Figure 6.2); such a broad zone of amphibolite facies is a consequence of the oblique cross section of volcanic crust exposed above the granite-greenstone contact dipping moderately away from a late (2699 Ma), crescentic granite

Table 6.9a. Metamorphic mineral assemblages in rocks of the Earngey–Costello area, Birch–Uchi greenstone belt (*modified from* Thurston 1985b).

Intermediate Metavolcanic Rocks	Mafic Metavolcanic Rocks	Ultramafic Intrusive Rocks
• albite + tremolite + biotite + chlorite ± quartz ± muscovite • albite + chlorite + carbonate ± biotite + quartz + epidote • albite + tremolite + epidote + biotite ± chlorite ± carbonate • albite + quartz + epidote + chlorite + biotite + muscovite + k–feldspar • albite + quartz + biotite + epidote ± carbonate • albite + tremolite + epidote ±carbonate + quartz • albite + chlorite + sericite + quartz • biotite + quartz + epidote + albite + carbonate + chlorite + tremolite • biotite + quartz + albite + epidote + stilpnomelane	• albite + tremolite + epidote + chlorite + biotite ± carbonate • albite + tremolite + quartz + epidote ± biotite ± chlorite ± carbonate • tremolite + albite + hornblende + epidote • chlorite + albite + carbonate + epidote • biotite + albite + quartz + carbonate ± chlorite ±epidote • albite + chlorite + epidote + quartz + carbonate	
• plagioclase + hornblende + quartz ± biotite ± epidote ± chlorite ± tremolite • hornblende + biotite + quartz + plagioclase + cummingtonite • plagioclase + hornblende + epidote + chlorite + carbonate + quartz ±biotite • plagioclase + hornblende + chlorite ±epidote ± quartz ±biotite • plagioclase + hornblende + quartz + chlorite + biotite ± epidote ± carbonate • plagioclase + hornblende + quartz + chlorite + epidote + biotite	• plagioclase + hornblende ± chlorite ± carbonate • hornblende + quartz + epidote + garnet ± carbonate • hornblende + plagioclase + biotite + chlorite + epidote ± quartz ± carbonate • plagioclase + hornblende + tremolite + biotite • plagioclase	
		All rocks contain plagioclase + epidote, with: • chlorite + tremolite-actinolite + sphene + sericite + carbonate + quartz • hornblende + sphene + chlorite + biotite + iron oxides + quartz • tremolite + carbonate + talc • actinolite + quartz + sphene + talc + anthophyllite
		• talc + carbonate + iron oxides + chlorite (peridotite) • talc + carbonate + iddingsite + iron oxides + chlorite • serpentine + chlorite + carbonate + tremolite • serpentine + talc + carbonate + chlorite

pluton in the batholith. This breadth of amphibolite facies metamorphism in the larger greenstone belts is not common, although similar, broad amphibolite facies domains have been reported in the vicinity of eastern Lake St. Joseph–Pashkokogan Lake (Goodwin 1965; Stott et al. 1987b) and in the Miminiska–Fort Hope greenstone belt (Thurston and Carter 1970; Thurston and Breaks 1978). More typical thermal aureoles are about 2 to 3 km wide; this can be seen, for example, in the northern and western granite-greenstone contacts of the Red Lake greenstone belt (Andrews et al. 1986) and in the Pickle Lake greenstone belt where the Ochig Lake pluton (*see* Figures 6.10 and 6.15) produced an amphibolitic aureole at least 3 km wide upon the enveloping basalts (Stott et al. 1989a). In western Lake St. Joseph, Clifford (1969) identified a typical basaltic assemblage of hornblende-almandine-quartz-plagioclase ± epidote within the thermal aureole of the Blackstone pluton. This contrasts with the regional greenschist facies metamorphism of most of the belt and the broader, high- to medium-grade thermal metamorphism and higher pressure conditions present in the adjacent English River Subprovince (Thurston and Breaks 1978; Chipera and Perkins 1988; *see* Breaks, this volume).

Conditions of low pressure metamorphism are recorded in eastern Lake St. Joseph where Goodwin (1965) and Thurston and Breaks (1978) documented medium-grade assemblages in wacke-pelite turbidites bearing staurolite + andalusite; similar conditions are recorded by Shklanka (1967) in the Bee Lake belt, where clastic sedimentary rocks locally contain andalusite with staurolite. Chipera (1985) summarized the conditions of metamorphism in the Lake St. Joseph area. The presence of andalusite limits the pressure conditions of metamorphism to less than 4 kbars (Holdaway 1971). The temperature was greater than 400°C owing to the common assemblage of chlorite-muscovite-biotite in sedimentary rocks; however, this

Table 6.9b. Common mineral assemblages in the metamorphic zones of the Miminiska Lake area (*modified from* Wallace 1981a).

Index Mineral (Zone)	Metamorphic Mineral Assemblages in Metapelites in the Miminiska Lake Area[1]	Comparable Metamorphic Grade (Winkler 1979)
chlorite	• sericite + plagioclase + chlorite	low
biotite	• biotite + sericite + plagioclase + chlorite • biotite + chlorite + plagioclase • biotite + plagioclase + sericite	low
garnet	• biotite + sericite + plagioclase + garnet (almandine) • biotite + chlorite + plagioclase + garnet (almandine) • biotite + plagioclase • chlorite + sericite + plagioclase + biotite + garnet (almandine)[2]	(almandine) low
staurolite	• staurolite + biotite + sericite + plagioclase + garnet (almandine) + chlorite • staurolite + biotite + plagioclase + andalusite + garnet (almandine) + chlorite • staurolite + biotite + plagioclase + andalusite + garnet (almandine) + sericite • biotite + plagioclase + garnet (almandine) + chlorite • biotite + sericite + plagioclase + chlorite[2]	(almandine + andalusite) medium

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[1] Quartz is present in all assemblages.

[2] Chlorite in the presence of sericite at conditions above biotite grade is believed to have formed by retrograde processes.

assemblage of the appropriate composition would break down since chlorite + muscovite + quartz may react to form cordierite + biotite + Al_2SiO_5 + H_2O at temperatures exceeding 500° to 550°C. Since cordierite has not been observed, this would limit conditions of metamorphism to less than about 550°C where the conditions of chlorite composition favoured cordierite formation (Winkler 1979). Staurolite, however, is more widespread in pelitic rocks in several of the belts and locally coexists with almandine garnet. The presence of staurolite indicates that temperature conditions probably reached 525° to 550°C at least locally (Hoschek 1969). The typical conditions of metamorphism in the greenstone belts, outside of thermal aureoles adjacent to the plutons, are thus constrained to pressures of less than 4 kbars and temperatures less than 550°C.

TIMING OF METAMORPHISM

Isotopic age determinations on several plutons, internal and external to the greenstone belts, provide general constraints on the timing of contact metamorphism and earlier regional metamorphism as shown on Figures 6.18a to 6.18d. In general, the most notable thermal aureoles around the late granitic plutons were imposed at approximately 2700 Ma, during and subsequent to the main episodes of regional, penetrative deformation across the subprovince and therefore constrain the principal periods of metamorphism recorded in the subprovince. This subprovince has not received much detailed analysis of its metamorphic history and awaits a more thorough documentation of the metamorphic sequence of events.

MINERAL DEPOSITS

Introduction

Exploration activity has occurred in the Uchi Subprovince during periodic bursts—since the 1920s and more recently in the 1980s when the belt attracted attention for its gold potential. At present there are 4 active gold mines in this subprovince shown on Figure 6.2, two in the Red Lake greenstone belt and recently opened mines in the Meen–Dempster and Pickle Lake greenstone belts. Other gold mines, principally in the Red Lake and Pickle Lake greenstone belts and to a lesser extent in the Birch–Uchi greenstone belt, have seen differing degrees of success and their records can be extracted from the annual *Report of Activities, Resident Geologists* reports published over the years by the Ontario Geological Survey.

A volcanogenic massive sulphide deposit in the Birch–Uchi greenstone belt experienced a brief but profitable life as the South Bay mine in the 1970s, but base metal exploration in this subprovince has received much less attention compared to exploration for gold. The former gold belt at Pickle Lake saw the geophysical discovery of a low-grade copper-nickel deposit, the Thierry deposit, which also had a brief history as a mine before closing in 1982. At present, there is continued exploration for gold, although not at the same intensity as in the 1980s, and a renewed interest in base metal exploration, especially in and southwest of the Birch–Uchi greenstone belt.

This section will review briefly some types of mineralization identified in the Uchi Subprovince. The deposits are shown in relation to the geology on Figure 6.2.

Deposit Types

GOLD DEPOSITS

The small mining communities clustered on the shores of Red Lake are the quintessential settlements founded on the discovery of gold in northern Ontario since the 1920s. However, Red Lake has had the good fortune of having 2 very successful mines still in operation after so many others in the same belt have closed. Over 16 million troy ounces of gold have been extracted from 17 mines in the Red Lake greenstone belt with over 7 million ounces coming from the Campbell Mine of Placer Dome Inc., which started production in 1949. This mine, at the end of 1990, had proven and probable reserves of over 3 million ounces of gold, with ore grading at 0.559 ounce Au per ton. The other mine still in operation—the Arthur W. White Mine of Dickenson Mines Limited—began gold production in 1948 and has produced over 2.7 million ounces of gold.

From the earliest investigations about the cause of gold mineralization in the Red Lake greenstone belt, it was commonly recognized that faults, fractures and shear zones owed their development to the stresses imposed upon the belt by the surrounding external granitic batholiths (Horwood 1948). Pirie (1981) noted a spatial association between most of the gold deposits in the eastern part of the Red Lake greenstone belt and a zone of pervasive iron carbonatization in the lower tholeiitic to komatiitic sequence, the Balmer assemblage. More recent investigations summarized by Andrews et al. (1986) suggested that the gold deposits, as complex as they are in detail, are essentially products of hydrothermal fluids introduced into ductile deformation zones at a late stage in the tectonic history of the belt and at about the time of plutonic emplacements (*see* Figure 6.17b). Some of the ductile deformation zones lie along or close to assemblage boundaries; the Flat Lake–Howey Bay deformation zone is an example of a zone within which lie discrete shear zones including the Austin shear zone, which hosts former gold mines (*see* Figure 6.17a).

In the Meen–Dempster greenstone belt, anomalous gold occurrences and the Golden Patricia Mine lie within deformation zones along or near the flanks of a strain domain shown in Figure 6.9 and described earlier. (The Golden Patricia Mine is owned by Lac Minerals Limited through their subsidiary Bond Gold Canada Inc.) The gold ore is in a narrow quartz sheet interpreted by Rodd and Hutchinson (1991) to be originally a chert unit rather than an epigenetic vein. But Rodd and Hutchinson (1991) recognized that the host rocks for the gold are sheared and the associated mineral assemblages and alteration zone are typical of the conditions observed elsewhere arising from late orogenic shear zone development with subsequent introduction of hydrothermal fluids. If the strain domain in which the shear zone-hosted gold occurs is a product of orogenic regional shortening during which the Dobie Lake pluton rotated and compressed the greenstone belt, then the mineralization would have concentrated during the late stages of the Kenoran Orogeny. Alternatively, the shear zone is a splay of the Bear Head fault zone, which cuts a 2696 million-year-old pluton, at a late stage when regional orogenic shortening was evolving to narrower zones of strain.

In the Pickle Lake greenstone belt, gold deposits (Barrett and Johnston 1948; Corking 1948; Cohoon 1986) are located in different strain domains shown in Figure 6.10. The Pickle Crow deposits are products of an early strain event that preceded the emplacement of the Hooker–Burkoski stock, which is coeval with the 2741 million-year-old Ochig Lake pluton. Yet, the Dona Lake Mine is localized in a folded, sheared and sulphidized iron formation unit within the strain aureole of the Ochig Lake pluton. One can interpret the shear zone as a product of contact strain and if confirmed, it would imply that both of these gold mineralization events in the Pickle Lake greenstone belt precede the main period of Kenoran orogenesis and thus predate the mineralization at Red Lake.

The structural setting for some gold deposits can be attributed directly to the emplacement of a felsic pluton as envisaged by Stott and Smith (1988) and illustrated in Figure 6.24. Within the contact strain aureole around a pluton, 5 types of gold-bearing shear zones can be identified. Auriferous shear zones of types C and D (*see* Figure 6.24) may be represented in the Pickle Lake greenstone belt:

Type A—set of shear zones bordering an uplifted and rotated wedge of supracrustal rock, for example, along the flank of the Canoe Lake stock in the western Wabigoon Subprovince

Type B—shear zone defining the outer margin of a contact strain aureole

Type C—shear zone focussed along a contrast in rock ductility, as in the vicinity of an iron formation unit in basalt, exemplified by the Dona Lake Mine of Placer Dome Inc.

Type D—shear zones in conjugate sets at small angles to the schistosity; this strain environment may correspond to regional orogenic shortening or to a strain aureole of a pluton; exemplified by at least some Pickle Crow gold deposits

Type E—shear zones, tangential to a strain aureole, may have been initiated during earlier orogenic shortening and locally reactivated

Table 6.10 summarizes the timing constraints for gold mineralization in the Red Lake and Pickle Lake greenstone belts, based on crosscutting relationships and structural criteria. The evidence for gold mineralization accompanying earlier magmatism and orogenic events (pre–2720 Ma) should be anticipated in the northern part of the Superior Province where there are broader terranes containing older Archean crust. Further discussion on this topic is given by Fyon, Breaks et al. (this volume).

VOLCANOGENIC MASSIVE SULPHIDE DEPOSITS

The South Bay Mine

The Uchi Subprovince contains scarce indications of volcanogenic massive sulphide deposits and only one—the South Bay Mine—has been profitably extracted. In 1969, Selco

Table 6.10. Time-constraining evidence for gold mineralization in the Uchi Subprovince.

RED LAKE[1] Gold Mineralization	Rock Type	Age (Ma)	Comments	PICKLE LAKE[2] Gold Mineralization	Rock Type	Age (Ma)	Comments
Flat Lake-Howey Bay deformation zone (2718–2699 Ma [min])	Dome Stock granodiorite	2718 (U-Pb zircon)	Host to auriferous quartz-vein deposits	Golden Patricia gold mineralization (2722 – <2696 (min) Ma) (ca. 2709 Ma)	Kawashe Lake stock (deformed by D_2 deformation in Meen–Dempster greenstone belt)	2722	The shear zone, containing gold mineralized unit, is interpreted to be formed during D_2 deformation, related to 2.7 Ga regional Kenoran orogenic shortening.
	Diorite dike near Madsen Mine	2699 (U-Pb titanite gives minimum age only)	Unaltered and undeformed dike that cuts across the altered and intensely deformed rocks of the Austin shear zone that hosts gold ore bodies; the diorite is one of a set of dikes that cut the ore bodies (Horwood 1945, Figure 43). This relationship can be further defined with tighter age determinations on granodiorite dikes that cut the ore bodies.		pluton (name?)	>2699	Pluton cut by the Bear Head fault; transcurrent displacement across the fault postdates the D_2 domain fabric hosting the gold.
				Dona Lake sheared and sulphidized iron formation (2741 Ma)	Ochig Lake tonalite pluton	2741 (approx. age)	Approximate age at which a contact strain and metamorphic aureole was produced; the mineralization lies within and carries the strain aureole fabric, the last penetrative deformation event in this region.
Cochenour–Gullrock Lake deformation zone (2720–2714 Ma)	McKenzie granodiorite	2720 (U-Pb zircon)	Sheared and highly altered host to auriferous quartz-vein deposits; foliations wrap around the body suggesting that most ductile deformation in the belt predates the stock. (This latter point is not a strong argument.)	Pickle Crow veins	—	pre–2741	These veins carry a penetrative fabric observed to be deflected by the local Dobie Lake suite of plutons.
	Quartz feldspar porphyry dike	2714 (U-Pb zircon)	Dikes cut across all southeast A.W. White mine ore bodies and yet show weak fabric suggesting emplacement during late-stage ductile deformation.				

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[1] Andrews et al. (1986)

[2] Stott et al. (1989a, 1989b)

Mining Corporation Limited discovered a stratiform copper-zinc-silver deposit in the uppermost rhyolitic sequence of the Confederation assemblage of the Birch–Uchi greenstone belt (Pollock et al. 1972), in that part formerly labelled Cycle III rhyolites by Thurston (1985a). This was a small deposit that produced, over its 10 year life span, 1.6 million tons of ore, averaging 11.06% Zn, 1.8% Cu and 2.12 ounces Ag per ton (Atkinson et al. 1990).

The ore lies, in association with an argillaceous chert horizon, above an endogenous rhyolite lava dome of quartz feldspar porphyry (H. Squair, unpublished Selco Mining Corporation report, 1973, as referenced by Urabe and Scott 1983) and under and within rhyolitic tuff breccia and lapilli tuff (Figure 6.26a). Hydrothermal alteration of the footwall quartz feldspar porphyry comprises 4 zones described by Urabe and Scott (1983, p.1868) as 1) sericite-buff carbonate; 2) incipient dark chlorite; 3) dark chlorite; and 4) bleached quartz feldspar porphyry (*see* Figures 6.26a and 6.26b). The deposit was described by Wan and Warburton (1979) as a set of stacked, irregular shaped ore lenses, of "massive fine pyrite, banded to interstitial chalcopyrite and finely banded to coarse massive sphalerite".

This westward-facing mine sequence is near the top of a fault-bounded rhyolitic block that was interpreted by Thurston et al. (1978) to be a caldera-like structure similar to calderas described by Smith and Bailey (1968). The ore deposit is typical of Archean volcanogenic massive sulphide deposits hosted by bimodal volcanic suites and interpreted by Cathles et al. (1983) to have formed in a volcanic arc or failed rifted arc environment.

The rare earth element (REE) patterns of the felsic volcanic host to this deposit are unusual, being comparatively unfractionated and with unusual light and heavy REE abundances (Thurston 1980b). (The recognition of anomalous REE patterns in mineralized calc-alkalic rhyolites of other greenstone belts has led various workers (Campbell et al. 1984) to investigate the degree of REE mobility and distinctive REE patterns (Lesher et al. 1986) as a potential exploration tool.)

Parts of the southern Birch–Uchi greenstone belt, mainly in the Confederation Lake area, have been reported to bear pre-deformational alteration in the form of epidotization, silicification, sericitization and chloritization (Fyon and Lane 1985). Occurrences of volcanogenic massive sulphide deposits have been reported elsewhere in the Confederation assemblage along strike from the South Bay Mine (Atkinson et al. 1990). Outside of the 2730 to 2740 million-year-old Confederation volcanic assemblage, there has been recent exploration of volcanogenic massive sulphide occurrences in the Woman assemblage east of Bamaji Lake and east of Pickle Lake.

MAGMATIC COPPER-NICKEL DEPOSITS

The potential for the discovery of magmatic copper-nickel deposits in the Uchi Subprovince has not been fully explored, although considerable exploration for such deposits was conducted by Umex Corporation after their discovery of the Thierry deposit, which had a short six-year history as an open pit and underground mine until its closure in 1982. This is the only copper-nickel deposit mined in the Uchi Subprovince.

The Thierry Mine

The Thierry copper-nickel mine in the Pickle Lake greenstone belt is a low-grade copper-nickel deposit in tabular intrusive gabbro, peridotite and dunite (altered to talc-carbonate schist) within a narrow arm of basalts wedged between the Williams Lake batholith and the Pickle Lake stock (*see* Figure 6.10). The host rocks are interpreted here to be part of an older assemblage, dominated by tholeiitic basalt in the northernmost part of the Pickle Lake greenstone belt. The rocks are strongly deformed, locally mylonitic and the ore is principally contained in a hornblende-biotite-chlorite schist (Verbeek et al. 1972; Sage and Breaks 1982, p.125; Figure 6.27). The sulphide ore is interpreted to be magmatic in origin (Patterson and Watkinson 1984a), but has been remobilized and metamorphosed to form several forms of ore association, notably as mylonite and breccia ore (Patterson and Watkinson 1984b). Observations by Stott et al. (1989a) of the sense of displacements in the host rock indicate that the strain aureole surrounding the Pickle Lake stock encompasses the Thierry deposit (*see* Figure 6.10). The remobilization and concentration of sulphide mineralization is attributable to the deformation imposed by the emplacement of the 2740 million-year-old Pickle Lake stock.

RARE ELEMENT PEGMATITES

The presence of subeconomic rare element pegmatite bodies (especially those containing lithium-bearing minerals) within the southern half of Uchi Subprovince has been known for many years (Sage and Breaks 1982; Kay and Stott 1985). Such bodies tend to concentrate near fundamental terrane boundaries in the Superior Province as described by several workers including Cerny and Meintzer (1988), Breaks (1989) and Breaks and Janes (1991) with useful references in each publication. Features of these pegmatites have been summarized by Cerny (1982) who supports a subdivision of the granitic pegmatites into 4 geological classes including the rare element class of mineralized, highly fractionated granites formed within volcanic and sedimentary rocks and metamorphosed to cordierite-amphibolite and upper greenschist facies in low pressure, Abukuma-type terranes. Such a metamorphic environment is preserved locally in the southern part of Uchi Subprovince adjacent to the higher grade English River paragneisses and granites. For example, staurolite-andalusite-bearing metasedimentary rocks and adjacent amphibolitized basalts in the vicinity of eastern Lake St. Joseph and Pashkokogan Lake (*see* Figure 6.25) contain muscovite-granite pegmatite bodies with anomalous, subeconomic concentrations of rare elements. These bodies are typically small, lenticular pods and dikes that conform to or crosscut the schistosity of the enveloping rocks. Kay and Stott (1985) confirmed Goodwin's (1965) assessment of these pegmatite dikes, which shows anomalous mineral chemistry of potassium feldspar (Figure 6.28) compared to analyses of fertile pegmatite

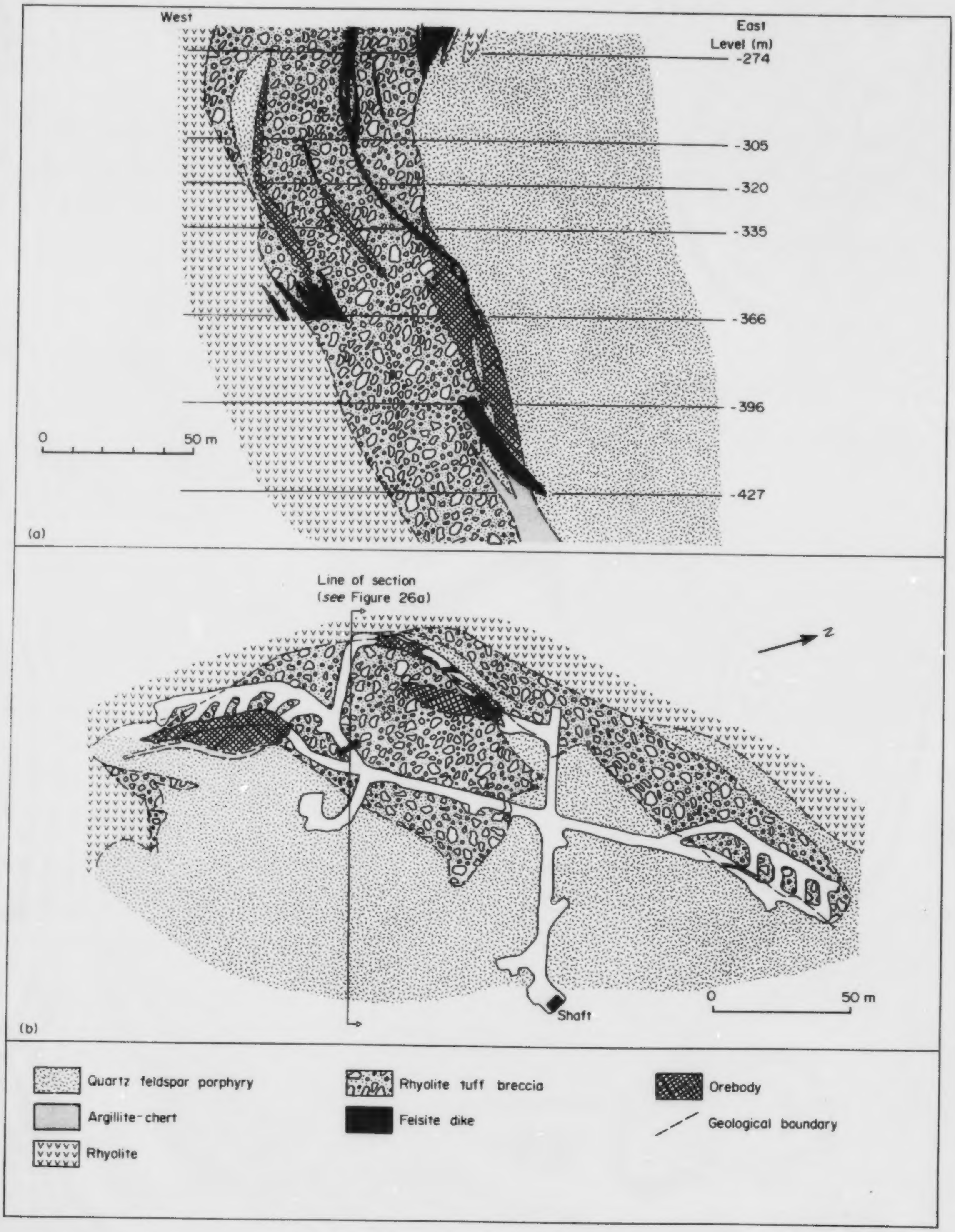
West
East
Level (m)
-274
-305
-320
-335
-366
-396
-427
0
50 m
(a)
Line of section
(*see* Figure 26a)
N
0
50 m
Shaft
(b)
Quartz feldspar porphyry
Argillite-chert
Rhyolite
Rhyolite tuff breccia
Felsite dike
Orebody
Geological boundary

Figure 6.26. a) A vertical section and, **b)** a level plan of the former South Bay Mine (copper-zinc) in the Confederation assemblage, Birch–Uchi greenstone belt (*see* Figure 6.5 for location). The relative position of the orebody to mappable units in the volcanic sequence is shown. (Figures *modified from* Urabe and Scott 1983; Thurston et al. 1978.)

bodies in southeastern Manitoba; these dikes also bear anomalous mineral assemblages (notably lithium-bearing, Table 6.11a) and rare element concentrations (Table 6.11b and Figure 6.29). A combination of appropriate metamorphic environment and proximity to the subprovince boundary restrict the spatial distribution of favourable localities of rare element pegmatite bodies to linear zones, recognized by Breaks and Osmani (1989) and described by Thurston, Osmani et al. (this volume).

OTHER UNUSUAL OR RARE MINERAL OCCURRENCES

Fluorite

Fluorite is reported as minor occurrences in pegmatite dikes near the Uchi–English River subprovince boundary (Dawson 1985, p.75); in the Springpole Lake area of the northern Birch–Uchi greenstone belt (Barron et al. 1989); and in the epizonal trondhjemitic North Bamaji pluton (Sage and Breaks 1982) where 0.10 to 0.20% purple fluorite, associated with molybdenum, occurs in coarse-grained quartz veinlets. The features of other industrial minerals in the Uchi Subprovince have been compiled and briefly described by Vos et al. (1987).

Molybdenite

Porphyry copper-molybdenum deposits are an uncommon feature in Archean terranes. However, 2 examples of molybdenum-bearing subvolcanic porphyry intrusions are noted within 35 km of each other in separate greenstone belts of the Uchi Subprovince. The first is within the margin of the Lang Lake belt, a 35 km long belt just north of the Meen–Dempster greenstone belt. Findlay and Ayres (1977) described the geological setting and the nature of the mineralization; evidence for the high-level subvolcanic emplacement of porphyry dikes adjacent to the Lang Lake belt is provided by the fine-grained to aphanitic groundmass of the dikes, the presence of chilled contacts and the lithologic complexity of several porphyritic units. The mineralization occurs within the various porphyry phases, especially the quartz-plagioclase porphyry, and the volcanic rocks that host the dikes. Pyrite, pyrrhotite, chalcopyrite and molybdenite occur as disseminations or in stringers and veins in fractures and shear zones. Associated with the mineralization is an alteration of the host rocks, which produced a mineral assemblage of epidote, quartz, chlorite, carbonate, sericite, albite, sphene and clay minerals.

The second example of an epizonal molybdenite occurrence is in a quartz vein system within the North Bamaji pluton (Sutherland 1978; Sage and Breaks 1982). The pluton is composed of fine-grained, slightly plagioclase-porphyritic trondhjemite that has been sheared and recrystallized. The

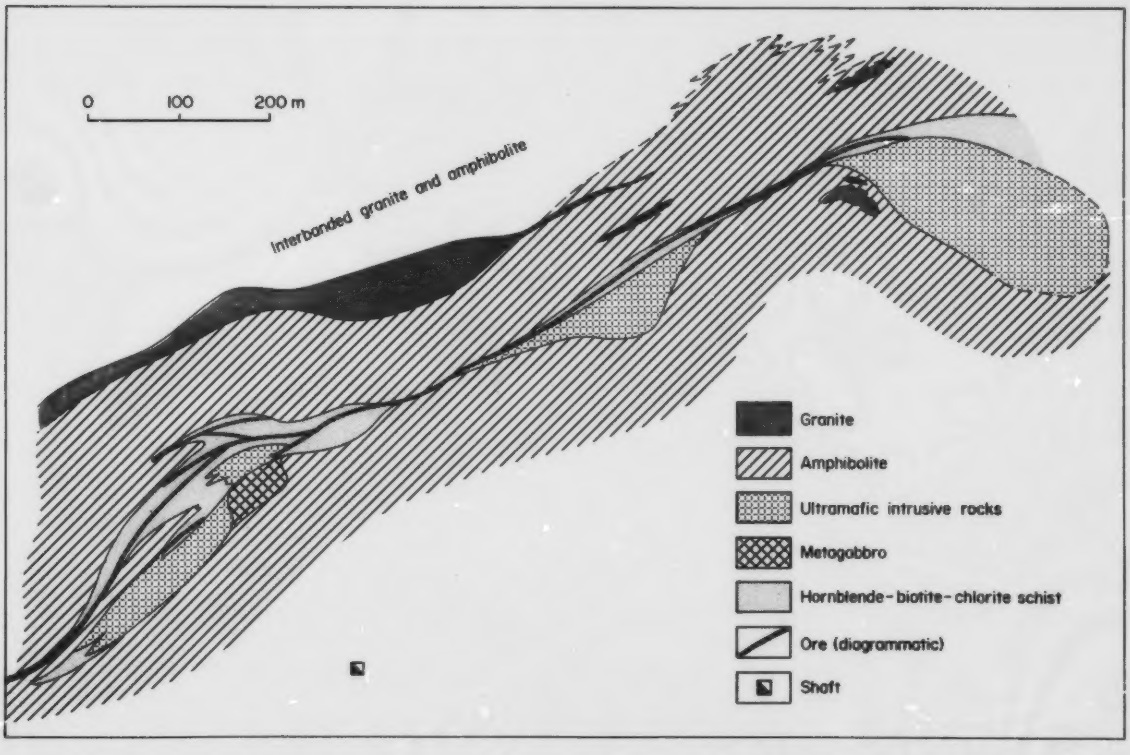
0
100
200 m
Interbanded granite and amphibolite
Granite
Amphibolite
Ultramafic intrusive rocks
Metagabbro
Hornblende-biotite-chlorite schist
Ore (diagrammatic)
Shaft

Figure 6.27. A geological level plan of the former Thierry Mine (copper-nickel; *see* Figure 6.10 for location) showing the ore restricted largely to the hornblende-biotite-chlorite schist, a shear zone in the mafic-ultramafic sill (*modified from* Sage and Breaks 1982; Verbeek et al. 1972).

Table 6.11a. Mineral assemblages of pegmatites observed in eastern Lake St. Joseph and east Pashkokogan Lake.

MINERAL	PEGMATITE East Pashkokogan Lake			Eastern Lake St. Joseph		
	1*	2*	3*	5*	7*	8*
Quartz	x	x	x	x	x	x
Plagioclase	x	x	x	x	x	x
Orthoclase	x	x	x	x	x	x
Spodumene		x				
Muscovite	x	x	x	x	x	x
Tourmaline	x	x	x	x	x	x
Garnet	x	x	x	x	x	x
Apatite		x	x			
Columbite-Tantalite?	x	x				

* Numbers indicate station location on Figure 6.29.

Table 6.11b. Rare element concentrations in pegmatite dikes.

Sample	Li ppm	Ta ppm	Nb ppm	Cs ppm	Be ppm	Ti %TiO_2	Bi ppm	Sn ppm
East Pashkokogan								
Sample 1	10	80	38	32	140	0.03	<0.1	87.1
Sample 2A	7100	26	31	115	140	0.03	0.2	136
Sample 2B	1260	19	36	33	100	0.05	0.1	133
Sample 2C	90	16	49	110	110	<0.01	0.1	126
Sample 3	16	23	26	130	150	0.05	<0.1	36.1
Eastern Lake St. Joseph								
Sample 4	5	23	36	50	15	0.05	2.7	4.6
Background in felsic intrusions[1]	40	20	10	2–6	20[2]	0.25	0.1	16.3[3]

[1] From Wedepohl (1978).
[2] Background concentration in pegmatites.
[3] Background concentration in Sn-bearing granites.

quartz veins hosting the molybdenum are emplaced in sheared and sericitized zones.

A third example of a molybdenum occurrence has been described by Beakhouse (1989) in granodiorite on the northwest margin of the Birch–Uchi greenstone belt. The disseminated and stringer molybdenite is concentrated along quartz-carbonate veins in zones of carbonate alteration.

Tectonic Assemblages, Structures and Future Exploration

Regional-scale correlations of assemblages encourage tracing potentially mineralized stratigraphic units over many 10s of kilometres and also permit predicting the correlation of stratigraphic units between greenstone belts. For example, stratiform, low-grade zinc occurrences are associated with dacitic pyroclastic units of the Woman assemblage in eastern Pickle Lake greenstone belt (P. Lewis, Minnova Inc., personal communication, 1989) and one might speculate that the units are correlative with coarser dacitic pyroclastic units and a subvolcanic porphyry intrusion on Keezhik Lake in the northern Miminiska–Fort Hope greenstone belt. The beginnings of a tectono-stratigraphic subdivision of Archean greenstone belts is an invaluable step towards defining the tectonic and depositional settings of the mineral deposits.

The authors can anticipate that future volcanogenic massive sulphide exploration will benefit from a clearer understanding as to how and when the greenstone belts and subprovinces were assembled; which stratigraphic units are potentially correlative, thereby tracing anomalous lithogeochemistry to larger deposits; and to answer questions such as whether (and why) there is a common episode of crustal growth in northwestern Ontario that is more fertile for volcanogenic massive sulphide deposits amongst all the subprovinces.

The authors recognize that a complex history has accompanied many of the fault zones in and adjacent to the greenstone belts, and not all regionally penetrative deformation is attributable to the *circa* 2700 million-year-old Kenoran Orogeny; broad parts of the belts can be shown to retain earlier orogenic histories, consistent with episodic accretionary growth, as we see from the record of growth of the Uchi–Sachigo superterrane (*see* Thurston, Osmani et al., this volume). Accordingly, gold may have been initially introduced at different episodes of crustal growth and it may record the local termination of brittle-ductile deformation (e.g., Pickle Crow Mine) or it may have been remobilized most advantageously during subsequent orogenic deformation (e.g., Golden Patricia Mine).

GEOPHYSICAL PATTERNS

Geophysical patterns have been identified which complement the geological features and provide some insight into the nature of the lithosphere underlying the Uchi Subprovince. The following is a summary of features and interpretations given by various workers from aeromagnetic and electromagnetic maps, two-dimensional gravity models, seismic traverses and paleomagnetic studies.

Aeromagnetic and Electromagnetic Patterns

Owing to the general sparcity of Proterozoic diabase dikes in the Uchi Subprovince, in distinct contrast to the Abitibi greenstone belt, for example, it is possible to extract meaningful patterns from the aeromagnetic maps of this region. McGrath and Hall (1969), for example, were able to show a close correlation between regional magnetic anomalies and major geological units in northwestern Ontario.

There are several features worthy of general comment, visible in the map of the total magnetic field of Ontario (*see* Map 2585, map case).

1. There is a close correspondence between major geological units and the aeromagnetic patterns. The lowest magnetic susceptibility within the subprovince is recorded generally within the greenstone belts, apart from the banded magnetite iron formation units and within some trondhjemitic plutons. The younger, granitic plutons and some tonalite gneiss terranes generally show a higher magnetic susceptibility.
2. Some of the late to posttectonic granite plutons are prominently displayed by their high magnetite content. Examples of the latter are found among the *circa* 2700 Ma, magnetite-rich, Williams granite-granodiorite suite (*see* Figure 6.3); this suite dominates the Berens River Subprovince.
3. One can integrate, with regional geological mapping, a qualitative analysis of the aeromagnetic patterns and susceptibility variations within granitoid batholiths to

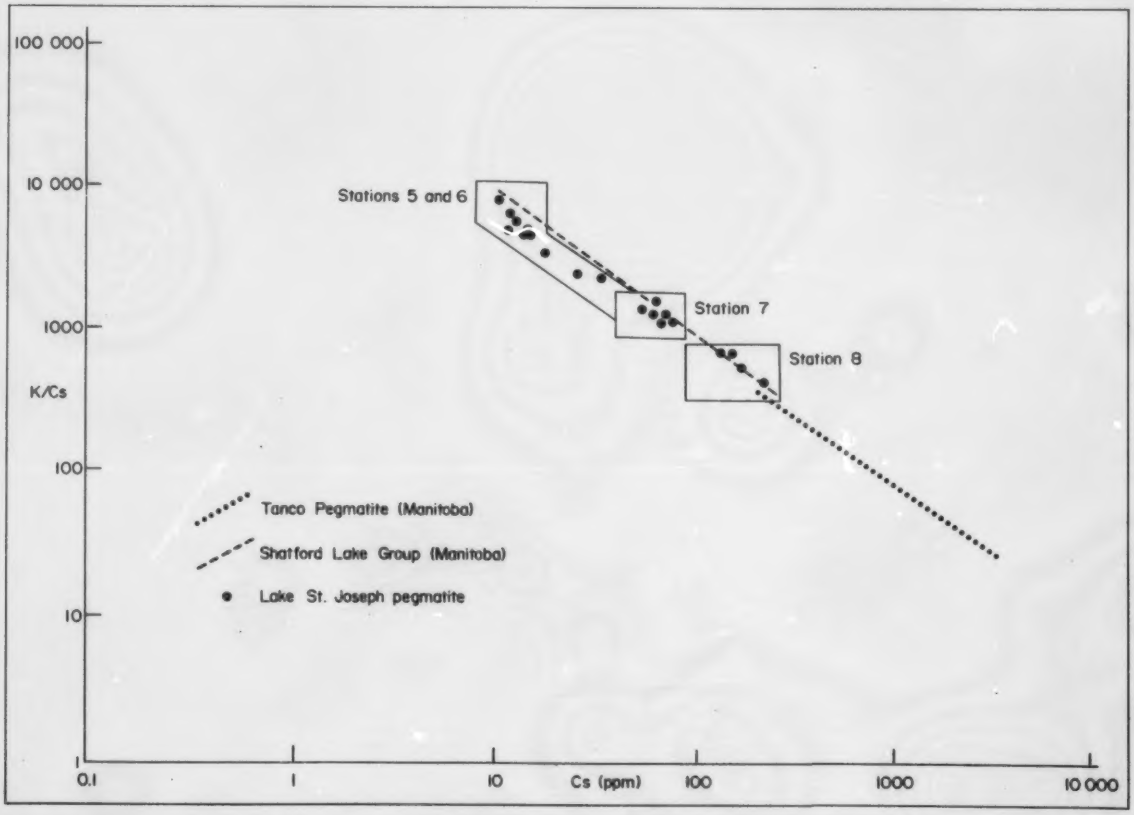

Figure 6.28. A comparison of K/Cs versus Cs for potassium feldspar samples in the pegmatites of Lake St. Joseph and southeastern Manitoba. Sample locations for the Lake St. Joseph pegmatites are shown in Figure 6.29.

distinguish bodies of foliated tonalite and gneiss from bodies of more massive, younger granite and granodiorite. This has permitted us to subdivide most of the batholithic complexes into their component parts to varying degrees of confidence, depending upon the available geological control. As a typical example, the aeromagnetic patterns of the Seach–Achapi granitoid complex, east of Pickle Lake greenstone belt (*see* Figure 6.2), closely correspond to the separate lithological units shown in Figures 6.3 and 6.10. The tonalite gneiss in the central core of the batholiths displays the most complex pattern of higher susceptibilities, contrasting with the susceptibility patterns in the adjacent granodiorite plutons.

4. Identification of faults and sense of displacement across them from aeromagnetic maps can be confirmed on the ground from locally developed kinematic

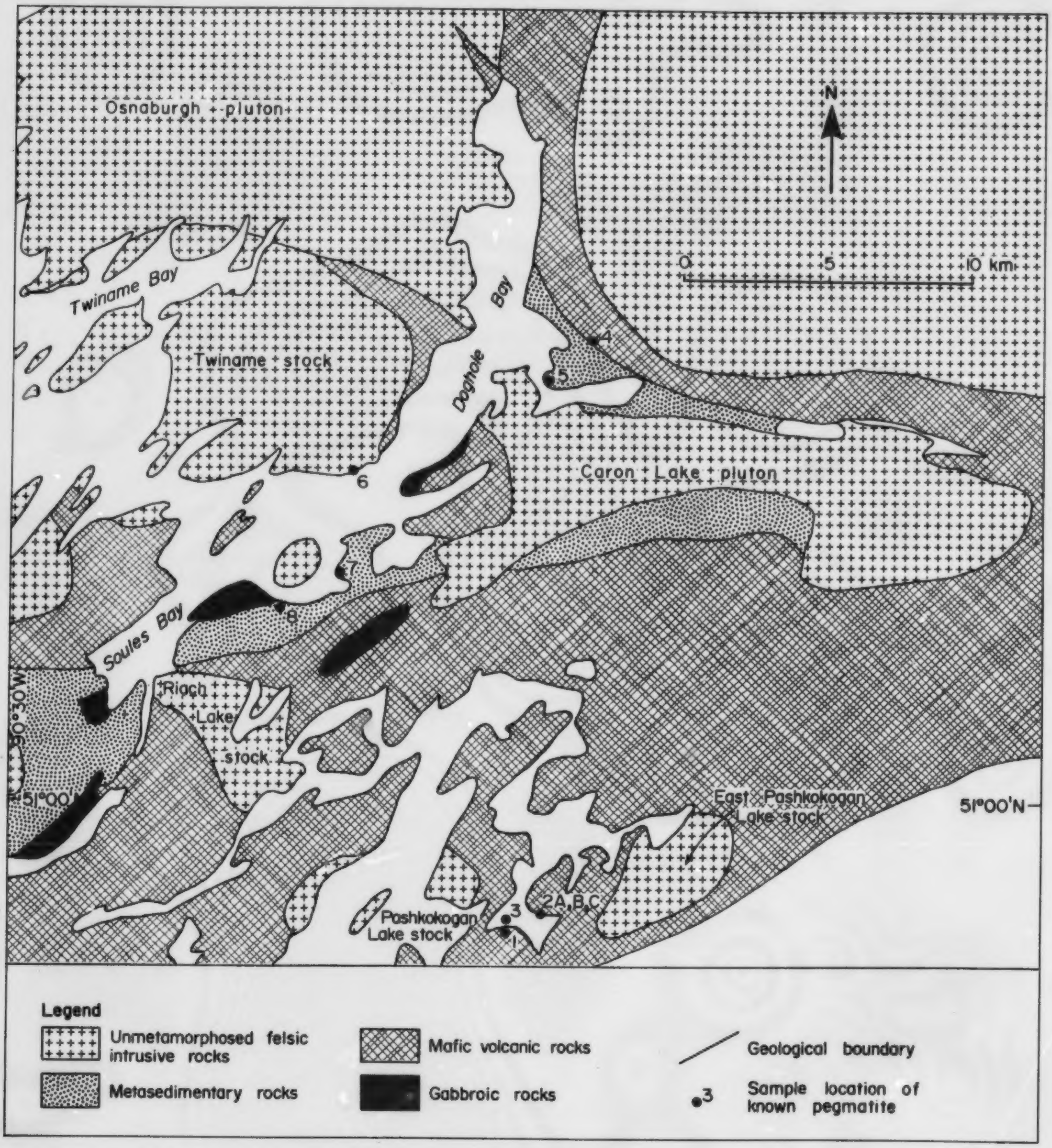
Osnaburgh pluton
N
0
5
10 km
Twiname Bay
Twiname stock
Doghole Bay
4
5
6
7
8
Caron Lake pluton
Soules Bay
Riach Lake stock
90°30'W
51°00'
East Pashkokogan Lake stock
51°00'N
Pashkokogan Lake stock
3
2A,B,C
1
Legend
Unmetamorphosed felsic intrusive rocks
Metasedimentary rocks
Mafic volcanic rocks
Gabbroic rocks
Geological boundary
3 Sample location of known pegmatite

Figure 6.29. The geology of eastern Lake St. Joseph showing location of pegmatite stocks and dikes from which samples were analyzed and compared with rare element-bearing pegmatites in Manitoba. One such comparison is shown in Figure 6.28.

criteria. Most prominent are the Bear Head fault zone, North Caribou-Totogan Lake shear zone and Sydney Lake–Lake St. Joseph fault zone (*see* Figure 6.3) described by Osmani and Stott (1988).

5. Aeromagnetic trends within the greenstone belts, notably defined by iron formation units, have been used in conjunction with stratigraphic markers and electromagnetic conductors to interpret the presence of obliquely juxtaposed tectonic assemblages. Two examples come from the Pickle Lake and northern Miminiska-Fort Hope greenstone belts. In the Pickle Lake greenstone belt, the combined geophysical features show that the unnamed volcanic domain to the north appears to be tectonically transected by the Pickle Crow assemblage, implying a fault juxtaposition (*see* Figure 6.10). An iron formation unit of the Pickle Crow assemblage may correlate with a similar unit in the northern Miminiska-Fort Hope greenstone belt, north of Keezhik Lake. The stratigraphy of the Keezhik Lake area is similar to and may correlate with the Pickle Crow assemblage. The iron formation unit stops abruptly against and at a high angle to another magnetite iron formation-bearing assemblage projecting from the northwest, possibly correlative with the McGruer assemblage of the North Caribou greenstone belt in Sachigo Subprovince. The oblique juxtaposition of geological units and geophysical trends may prove to be widely useful in exposing the presence of suspected fault boundaries between tectonic assemblages, especially in terrains covered by extensive overburden.

Gravity Patterns

The shape and thickness of greenstone belts have intrigued geophysical investigators. Grant et al. (1965) determined the maximum thickness of the Red Lake greenstone belt, from gravity measurements taken along 2 traverses, to be no more than about 4.7 km. This compares with the results of a gravity survey encompassing both the Red Lake and Birch–Uchi greenstone belts by Gupta and others (Gupta et al. 1982; Gupta and Wadge 1986). By using data from 5180 gravity stations, they concluded from two-dimensional gravity modelling that the depth-extent of the Red Lake greenstone belt averages 4 km and the Birch–Uchi greenstone belt, 3.5 to 4 km, with the narrow belt of supracrustal rocks connecting these belts averaging about 2.3 km depth. The volcanic and sedimentary rocks in this area range in depth-extent from 0.5 to 5.5 km.

Similar depths of greenstone belts have been independently demonstrated in other subprovinces from gravity modelling; for example, Dusanowskyj and West (1976) concluded that the Sturgeon Lake greenstone belt of the Wabigoon Subprovince is broad and flat at its base, 3 km deep; Gupta and Osmani (1990) modelled part of the Abitibi greenstone belt and found that the depth-extent ranges from 3 to 12 km (Gupta and Osmani, unpublished data, 1991). In the Manitoba part of the Uchi Subprovince, Brisbin (1971) determined depth estimates of the Rice Lake–Wallace Lake greenstone belt to range from approximately 3 to 5.5 km.

Gupta and Osmani (Ontario Geological Survey, personal communication, 1991) have determined that only in the youngest volcanic belt—the Abitibi—is there significant variation in depth to its base; most gravity transects of other greenstone belts show a thin, supracrustal "scum" only 1 to 6 km from surface to a comparatively flat base. What is the mechanism to produce such a shallow vertical extent to these greenstone belts, which appear surprisingly rootless? In western Uchi Subprovince, Gupta et al. (1982) proposed that the model best fitting the gravity and geological data supposed 2 mechanisms: large-scale thrusting and stacking of volcanic strata, and magmatic stoping to form granitic magma chambers during late volcanism and during subsequent orogenic plutonism. These suppositions are more recently encouraged by geochronological evidence of repeat-stacking of volcanic sequences and stacking out of stratigraphic order in the Confederation assemblage (Corfu and Stott 1989; Noble 1989). Furthermore, the central gravity low on the profile across the Birch–Uchi greenstone belt (Figures 6.30a and 6.30b) is consistent with the hypothesis for a resurgent caldera in a northward-striking graben (Thurston 1985a).

Figure 6.30a demonstrates that the residual gravity anomalies, determined from graphical smoothing procedures (Gupta and Ramani 1980), show a close correspondence with the greenstone belt geology as we noted above. These anomalies are separated from the regional component of the Bouguer gravity, which displays an eastward-trending gravity low (*see* Figure 6.30b) coinciding with the Uchi Subprovince and contrasting with a positive regional anomaly along the English River Subprovince (Gupta and Barlow 1984); this feature is also evident on a map of the vertical gradient of the Bouguer gravity anomaly (Goodacre et al. 1989); the pattern in the Uchi Subprovince was attributed to undulations of intralithospheric discontinuities (Gupta and Ramani 1980). Such a gravity anomaly pair has also been documented in southern Africa across a comparable structural boundary between the Limpopo belt and the Kaapvaal craton (Eminike 1986; Kleywegt 1988). Thomas (1989) suggested that the pair of regional positive-negative anomalies across the English River–Uchi subprovince boundary is remarkably similar to anomaly pairs along Proterozoic structural boundaries in the Canadian Shield, such as the Circum–Superior fold belt. The implication is that, like the Proterozoic boundaries, the English River–Uchi boundary is a zone of collision between 2 terranes. However, Thomas (personal communication, 1989) interpreted such a geophysical anomaly to mean that the Uchi–Sachigo superterrane was subducting under the Winnipeg River terrane at the time of collision. But the geological evidence of successive continental growth southward across the Uchi Subprovince plus the abundance of 2700 million-year-old granite plutons in the Uchi Subprovince and dominating a batholith chain farther north would appear to favour a

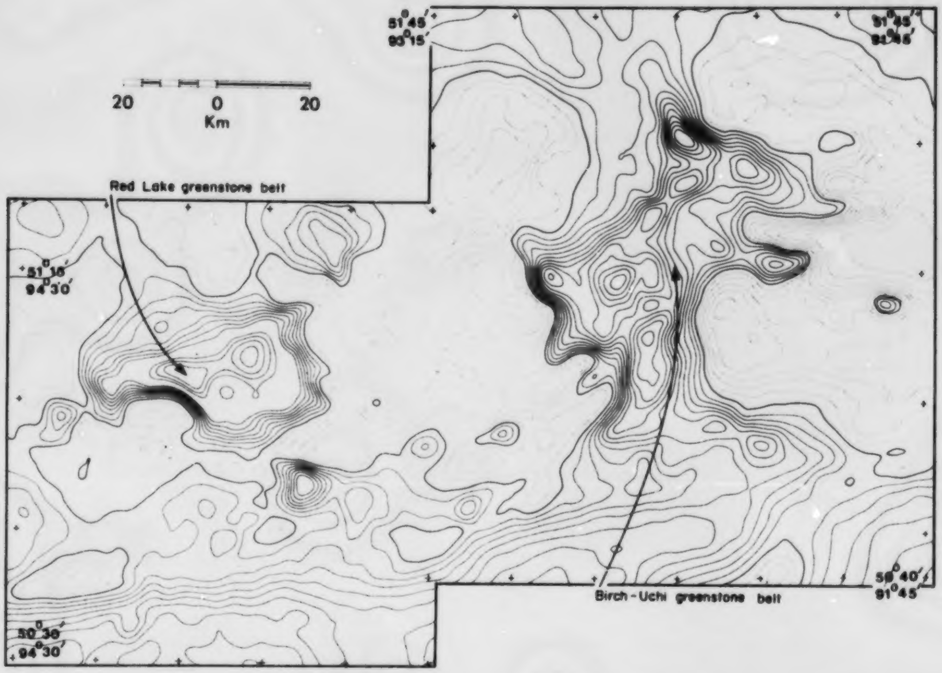
51°45′
93°15′
51°45′
91°45′
20
0
20
Km
Red Lake greenstone belt
51°15′
94°30′
50°40′
91°45′
Birch - Uchi greenstone belt
50°30′
94°30′

Figure 6.30a. Graphically separated residual component of the Bouguer gravity map in the region of Red Lake–Birch–Uchi greenstone belts. Positive contoured areas are centred on the 2 greenstone belts and shown with dark solid lines. Negative contours are shown by light, stipled lines. The zero contour is marked by a bold, solid line. Contour interval is 2 mgal (*modified from* Gupta and Wadge 1986).

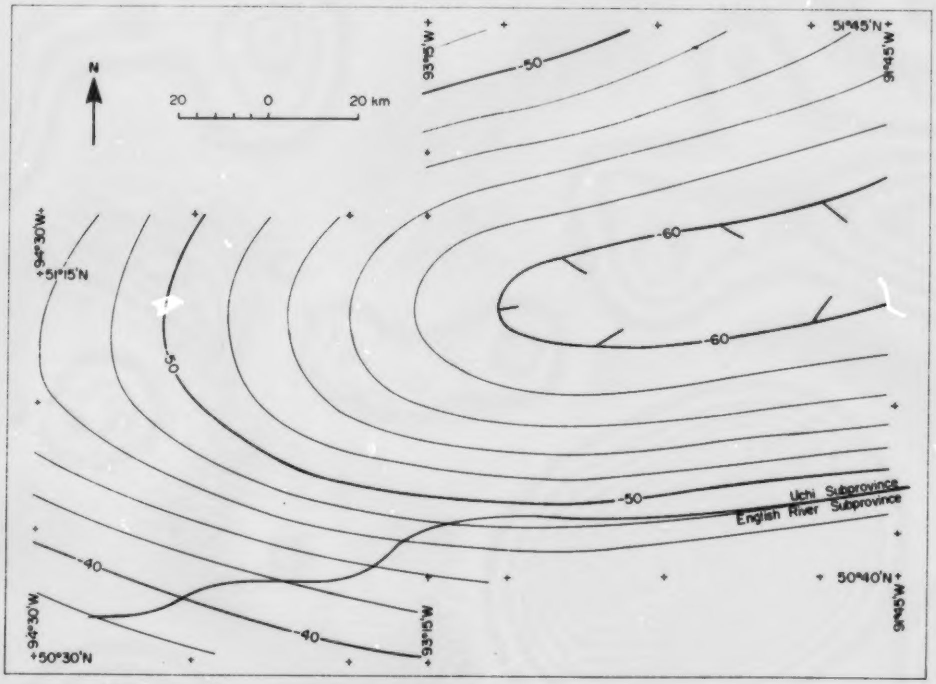
N
20
0
20 km
93°15'W
51°45'N
91°45'W
-50
94°30'W
51°15'N
-60
-60
-50
-50
Uchi Subprovince
English River Subprovince
-40
-40
50°40'N
91°45'W
94°30'W
50°30'N
93°15'W

Figure 6.30b. Graphically separated regional component of the Bouguer gravity map in the Red Lake–Birch–Uchi greenstone belt region showing a regional gravity low within the Uchi Subprovince and trending parallel to the boundary with the English River Subprovince. Contour interval is 2 mgal (*modified from* Gupta and Wadge 1986).

northward subduction of lithosphere under the Uchi–Sachigo superterrane at the time of the Kenoran Orogeny.

Seismic Patterns

Differences in geophysical characteristics between subprovinces can also be obtained from seismic reflection and refraction. Investigators at the University of Manitoba (Hall and Hajnal 1969; Hall and Brisbin 1982) showed that there are significant differences in the depths to crustal discontinuities and to the seismic crust-mantle discontinuity (the Moho) as one traverses the various subprovinces. The depth to the Moho (Figure 6.31a) increases, as one proceeds northward, from a minimum of about 30 km under the English River metasedimentary gneisses to 34 km under the Uchi granite-greenstone crust and 36 km under the granitoid rocks of the Berens River Subprovince (Hall and Hajnal 1969; Hall 1971). This variation contrasts with the depth variation to an intermediate discontinuity of 18 to 22 km as shown in Figure 6.31b. This intermediate discontinuity, or change in seismic velocity, was confirmed by James and Clarke (1990). It is noteworthy that the crust is thinnest in this seismic profile where it coincides with the Sydney Lake–Lake St. Joseph Fault along the Uchi–English River boundary.

The upper mantle, underlying the western Superior Province, has been investigated by Silver and Chan (1988) who analyzed the velocity of seismic shear waves passing through it. They found that SKS shear waves (one of a family of seismic waves that pass through the Earth's outer core and mantle) polarize or split into 2 orthogonal directions each with a different velocity. This shear wave splitting reflects the strain-induced alignment of anisotropic crystals of olivine in the mantle. The resulting velocity split into slow-velocity and fast-velocity directions, governed by the crystalline anisotropy, can be used to estimate the principal axes of strain, the depth-extent of anisotropy and its magnitude. The investigations of Silver and Chan (1988) encouraged them to advocate that the anisotropy measured is a "fossil" anisotropy developed in the upper mantle during the late Archean. This implied that we could use seismology to discover and analyze the characteristics of an Archean mantle still preserved as a root under the Canadian shield. Indeed, Silver and Chan concluded that at least part of the Superior Province is underlain by a 200 km thick keel of uppermost mantle, which has remained stable since Archean time. Such lithospheric strength may account for the size of the Superior Province, preserved as an Archean craton through time and subsequent break up of continents. The fast-velocity direction, recorded below the Uchi–English River subprovince boundary by Silver and Chan (1988), closely follows the trend of not only the subprovince boundary but also the dominantly east-northeast trend of the tectonic assemblages in the Uchi Subprovince. One may infer that the mantle anisotropy reflects the same lithospheric accretion and deformation processes that produced the belt-like subprovinces during the Kenoran Orogeny about 2700 Ma. Such a pattern of anisotropy merits comparison with the fossil anisotropy of the upper mantle recorded below Phanerozoic orogens where, in each case, the fast-velocity direction is parallel to the regional trend of the mountain belts and reflects mantle strain during orogenesis (Silver and Chan 1988; Shearer 1988; Vauchez and Nicolas 1991).

Silver and Chan's (1988) analysis of the results from a seismograph station near Red Lake indicated that the upper mantle of the Superior Province is characterized by the fastest seismic velocities and the deepest anisotropic keel in North America. This discovery confirmed the presence of a high velocity mantle root as determined by Grand (1987). This was followed by an experiment involving an array of portable three-component seismometers set out along a 1500 km traverse, including a section in Ontario from the Berens River–Sachigo subprovince boundary to the Quetico Subprovince. The results, reported by James and Clarke (1990), confirmed the presence of anisotropy in the upper mantle and identified evidence of discontinuities at depths of 80 to 200 km.

Since the upper mantle mineral fabric evidently preserves an Archean deformation, Silver and Chan (1988) recognized that the upper mantle beneath the Red Lake station would have had to have been sufficiently cold to preserve the strained crystal fabric. These authors advocated that this stability reflects a mechanical strength that probably arose from strain hardening during orogenic deformation. This would also imply that the mantle root did not grow through post–Archean underplating, confirming the conclusions of Bell and Blenkinsop (1987).

Paleomagnetic Patterns

Paleomagnetic studies have only recently been ventured in the Uchi Subprovince, notably by Hale and Lloyd (1989, 1990) in central Uchi Subprovince and by Costanzo-Alvarez and Dunlop (1990) in the vicinity of Red Lake greenstone belt. From selected granitic and gabbroic plutons in both areas, these workers obtained stable magnetizations that are inconsistent with the presently published Archean Aparent Polar Wander (APW) Path for the Superior Province (Irving 1979). Both sets of workers have advocated a revision to the APW Path (e.g., Hale and Lloyd 1990; Hale 1991). As additional, stable magnetizations become available from late Archean plutons in different subprovinces, one can anticipate that a new APW Path will provide useful constraints on the timing and nature of post–2700 Ma intracratonic crustal movement. However, the presently available determinations on late tectonic plutons (2.7 Ga) in both Uchi and Abitibi subprovinces show that the plutons bear very similar paleomagnetic poles and accordingly indicate that the amount of intracratonic movement across the Superior Province after about 2.7 Ga was rather limited (Hale 1991).

Block rotation of crust between late, parallel faults has been proposed on a regional scale, especially in the Sachigo Subprovince, where clockwise rotations are evident from a qualitative analysis of asymmetric, sygmoidal aeromagnetic patterns (Osmani and Stott 1988). Such clockwise rotation on a local scale has been suggested independently by Hale and Lloyd (1989, 1990) in the central part of Uchi

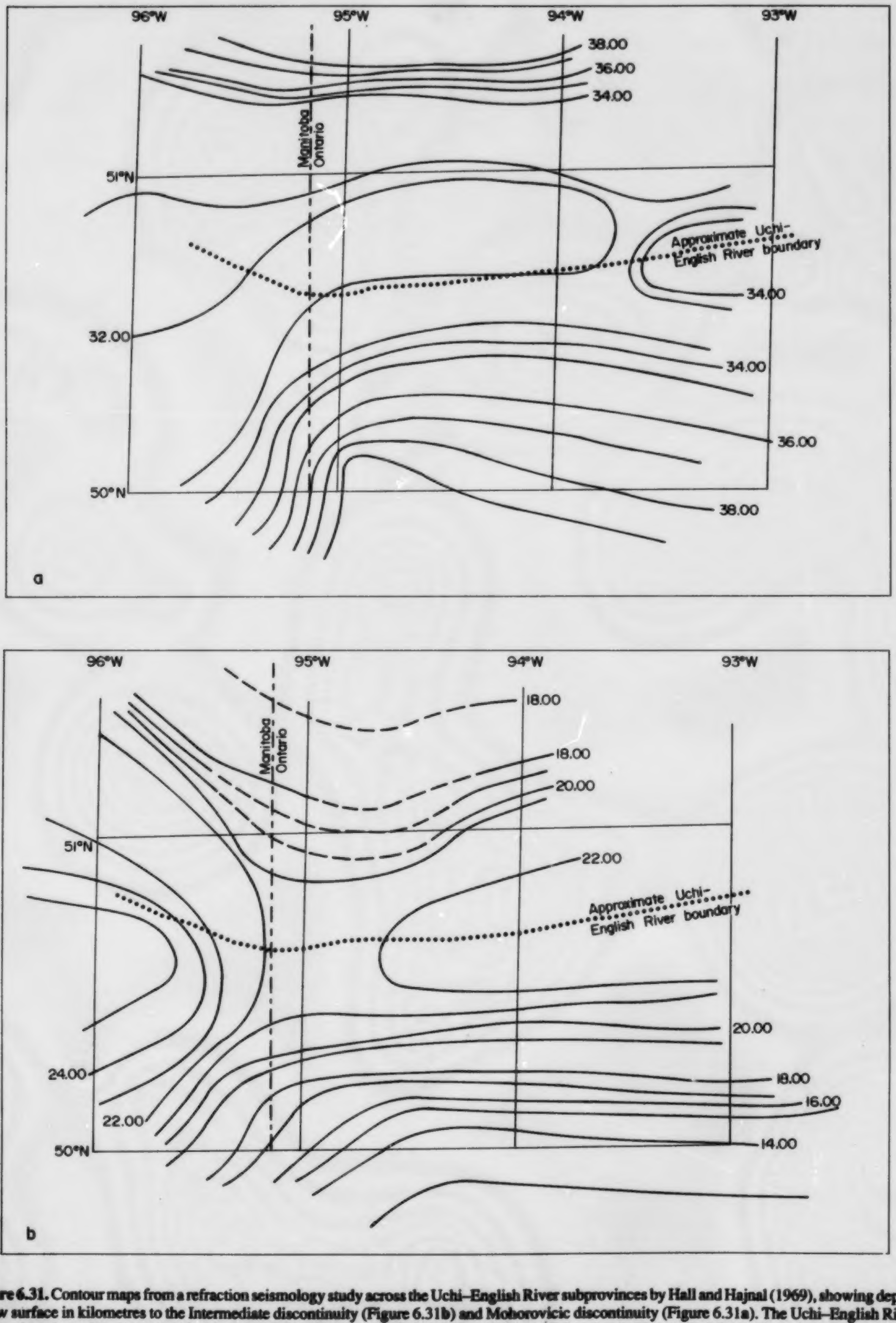
96°W
95°W
94°W
93°W
38.00
36.00
34.00
Manitoba
Ontario
51°N
Approximate Uchi-English River boundary
34.00
32.00
34.00
36.00
50°N
38.00
a
96°W
95°W
94°W
93°W
18.00
18.00
20.00
Manitoba
Ontario
51°N
22.00
Approximate Uchi-English River boundary
20.00
24.00
18.00
16.00
22.00
50°N
14.00
b

Figure 6.31. Contour maps from a refraction seismology study across the Uchi–English River subprovinces by Hall and Hajnal (1969), showing depths below surface in kilometres to the Intermediate discontinuity (Figure 6.31b) and Mohorovicic discontinuity (Figure 6.31a). The Uchi–English River subprovince boundary is marked by the rise of the Mohorovicic discontinuity to about 32 km depth. The dashed lines shown in part **b** represent interpreted contour values. (Figures *modified from* Hall and Hajnal 1969.)

Subprovince by comparing the orientations of stable remanent magnetizations of plutons within and outside of the rotated block. With available age determinations of plutons (F. Corfu, unpublished data; Corfu and Stott 1989), Hale and Lloyd concluded that, at some stage in the interval between 2747 and 2722 Ma, a block of the Meen–Dempster greenstone belt, containing the Kawashe Lake stock, rotated clockwise relative to the Dobie Lake pluton, north of the belt. This block rotation compares closely with the conclusions derived from geological and structural arguments, as presented for the Meen–Dempster greenstone belt in the section entitled Regional Penetrative Structures.

SEQUENCE OF EVENTS IN THE UCHI SUBPROVINCE

Each of the greenstone belts has recorded a sequence of magmatic and structural events, the highlights of which are described in the sections entitled Intrusions and Tectonic Elements—Structures and Metamorphism and are summarized in Figures 6.17b and 6.18a to 6.18d for most of the belts. The authors have reviewed these events and recognized patterns that developed across time and space in the creation of the Uchi Subprovince. This section summarizes these patterns and the sequence of events.

Patterns of Secular and Spatial Progressions

1. There are secular and spatial progressions in lithologies across the subprovince. The older assemblages, especially the Balmer and the northern half of the Pickle Lake greenstone belt, are dominantly composed of oceanic tholeiitic and komatiitic basalt flows. These contrast with younger assemblages, consisting of greater volumes of successfully preserved calc-alkalic volcanic rocks, with more voluminous rhyolite in some assemblages and greater andesite-dacite pyroclastic deposits in others. Coupled with this is the observation that the change in volcanic composition occurs southward and eastward across the subprovince, culminating in the more voluminous calc-alkalic deposits of the 2723 to 2713 million-year-old St. Joseph assemblage. There are also local secular progressions from calc-alkalic to alkalic volcanism in the 2781 million-year-old Bamaji assemblage at Bamaji Lake (Wallace 1985) and in the *circa* 2740 million-year-old Confederation assemblage plus intrusions at Springpole Lake (Barron, in press) in the northern Birch-Uchi greenstone belt.

Sequences of clastic sedimentary rocks lie unconformably upon the assemblages and in some cases overlap the volcanic assemblage boundaries. These sequences developed in late-forming basins and record an episode of exhumation and erosion of continental crust during the Kenoran Orogeny.

A secular progression in plutonic compositions is evident: older tonalitic gneisses (e.g., Pembina Tonalite Gneiss of the Lake St. Joseph batholith) and synvolcanic trondhjemitic plutons (e.g., elongate plutons within the Confederation assemblage of the Birch–Uchi greenstone belt) are followed by more potassic plutonism with increasing granite to granodiorite compositions amongst the younger 2713 to 2693 million-year-old plutons. There is no general spatial progression in the ages of plutonic rocks, although in some cases, ages of felsic volcanism in the southern half of the subprovince correspond with comparable ages of plutonism in the northern half. This correspondence of volcanism and plutonism in different parts of the subprovince is a critical feature that sets limits to the probable way in which the subprovince was constructed, as interpreted later.

2. There are secular and spatial progressions in structural style across the subprovince. The "homofacing" panels of strata in the northern half of the subprovince are in some cases repeated by stacking, as in the Confederation assemblage within the Birch–Uchi and Meen–Dempster greenstone belts; in other cases, they are stacked out of normal stratigraphic order, as in the eastern part of the Red Lake greenstone belt and in western Lake St. Joseph. This style, reported also by Corfu and Ayres (1991) for the Favourable Lake greenstone belt (Sachigo Subprovince), is inferred to be a product of thrusting and is superseded by orogen-parallel folding, especially near the southern margin of the belt. Moderately plunging, stretching lineations in the wider belts (viewed as orogen-parallel lineations) are superseded by downdip lineations associated with folding and thrusting near the southern margin of the belt (comparable to fabrics observed along the southern margin of the Wabigoon Subprovince by Williams (1987) and within the strain aureoles of plutons. Penetrative fabric development across the subprovince was succeeded by wrench faulting. The Sydney Lake–Lake St. Joseph Fault began as a thrust fault, transecting volcanic strata, and terminated as a ductile to brittle-ductile wrench fault, reflecting the final stage of north-northwest-directed, regional shortening and telescoping of the subprovince.

Of note is an anisotropic upper mantle mineral fabric, interpreted by Silver and Chan (1988), which preserves an orogen-parallel extension direction and compares favourably with the regional trend of greenstone belts and the long axis of elliptical plutons. Such a linear, belt-like trend is similar to the form of Proterozoic and Phanerozoic orogens.

Sequence of Post–2740 Ma Events

Although older deformation (pre–2740 Ma) is recorded locally, for example in the Pickle Lake greenstone belt, the most readily documented sequence of deformation events occurred subsequently. This sequence of deformation events has been ascertained in the major greenstone belts (e.g., Figures 6.18a to 6.18d) and details summarizing structural style and events for 6 belts were given in the section entitled Regional Penetrative Structures. From these features, a general sequence of structural events can be proposed, beginning with the formation of assemblage boundaries in the southern part of the Red Lake greenstone belt, which predates the main phase of the Kenoran Orogeny, that follows 2713 Ma volcanism. The sequence involves overlapping episodes of volcanism, thrust stacking and compression of strata, and tectonic juxtaposition of assemblages, crustal exhumation, erosion, folding and wrench faulting.

THRUSTING AND COMPRESSION

There is an episodic period of thrust stacking and folding of volcanic and sedimentary strata. The initial episodes in the Red Lake greenstone belt culminated in the juxtaposition of the Balmer and Confederation assemblages along a tectonic boundary before 2718 Ma, the age of the crosscutting Dome stock. Similarly, in the Meen–Dempster greenstone belt, tectonic imbrication of volcanic strata occurred before 2722 Ma, the age of the Kawashe Lake stock, which intruded the Kawashe gabbro, stitching the tectonic boundary between the Woman and Confederation assemblages. These thrust stacking events were synchronous with St. Joseph assemblage volcanism.

A subsequent episode of thrust stacking postdates the youngest cycle of the St. Joseph assemblage, that is, after 2713 Ma. This episode involved thrust repetition of the Confederation and St. Joseph assemblages, observed in western Lake St. Joseph. It probably marks the commencement of the main stage of the Kenoran Orogeny. This northward thrusting of strata formed stacks of volcanic and sedimentary sequences in the eastern half of the Birch–Uchi greenstone belt, where age determinations (Noble et al. 1989) show that the Confederation volcanic cycles are stacked out of normal stratigraphic sequence, that is, younger strata lie under and to the southeast of older Confederation strata. This lobe of tectonically interleaved volcanic and sedimentary units, closing to the north, appears to characterize the eastern half of the belt. This lobe is part of the postulated "South Uchi parautochthon" shown in Figure 6.32. The evidence of truncated volcanic bedding southeast of Lake St. Joseph (Goodwin 1965) indirectly confirms a

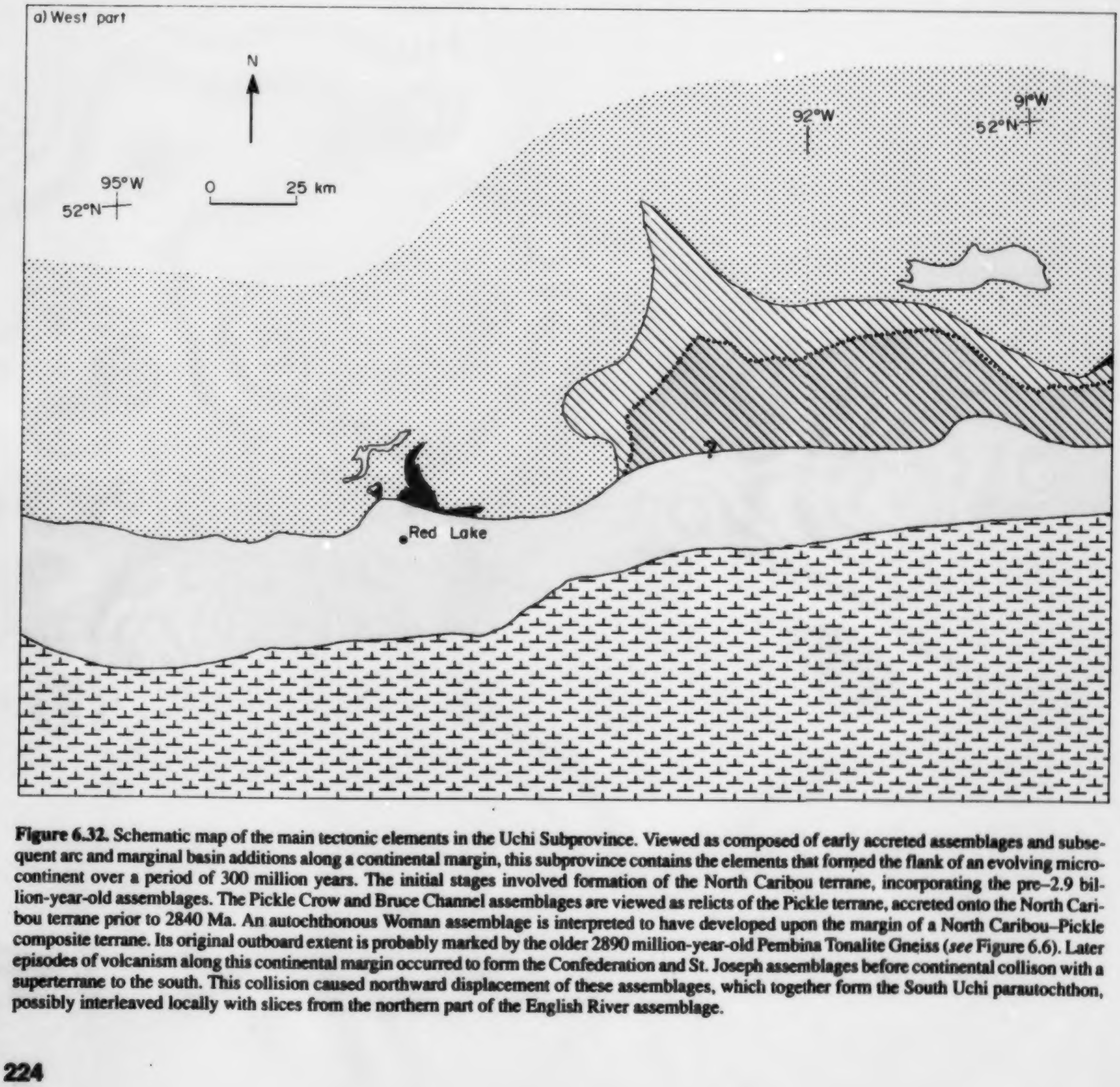

Figure 6.32. Schematic map of the main tectonic elements in the Uchi Subprovince. Viewed as composed of early accreted assemblages and subsequent arc and marginal basin additions along a continental margin, this subprovince contains the elements that formed the flank of an evolving microcontinent over a period of 300 million years. The initial stages involved formation of the North Caribou terrane, incorporating the pre–2.9 billion-year-old assemblages. The Pickle Crow and Bruce Channel assemblages are viewed as relicts of the Pickle terrane, accreted onto the North Caribou terrane prior to 2840 Ma. An autochthonous Woman assemblage is interpreted to have developed upon the margin of a North Caribou–Pickle composite terrane. Its original outboard extent is probably marked by the older 2890 million-year-old Pembina Tonalite Gneiss (*see* Figure 6.6). Later episodes of volcanism along this continental margin occurred to form the Confederation and St. Joseph assemblages before continental collison with a superterrane to the south. This collision caused northward displacement of these assemblages, which together form the South Uchi parautochthon, possibly interleaved locally with slices from the northern part of the English River assemblage.

stage of northward thrusting along the Uchi–English River subprovince boundary, removing part of the St. Joseph assemblage from the southern margin of the subprovince.

EXHUMATION

Greenstone belts and synvolcanic plutons were exhumed subsequent to at least the initial stages of thrust stacking of the volcanic strata. The eroded detritus from these rocks collected in syntectonic basins to form sedimentary sequences. These younger sequences lie unconformably upon the volcanic rocks, as in the Meen–Dempster greenstone belt, where the sedimentary Billett assemblage containing trondhjemitic clasts of the nearby Dobie Lake pluton forms an overlapping sequence upon the Woman–Confederation assemblage boundary and is itself folded during northward-directed regional shortening. Another example is the Eagle Island assemblage in the Lake St. Joseph greenstone belt where a basal conglomerate contains clasts of the underlying St. Joseph assemblage and pink arkose is presumed to be derived from the erosion of a granitic pluton. This assemblage unconformably overlies the fault boundary separating the St. Joseph and Confederation assemblages.

FOLDING

The above processes of thrusting and exhumation are accompanied and followed by folding, locally concentrated close to the southern boundary, as in the Bee Lake belt and Lake St. Joseph greenstone belt and in the syntectonic sedimentary sequences; folding in the Lake St. Joseph area followed 2713 Ma volcanism and preceded the emplacement of the 2702 million-year-old Blackstone pluton. The crust in this area, initially thickened by thrust stacking and then folded, reflects a regional progression in the style of deformation from "thin-skin" to "thick-skin" processes as observed in Phanerozoic orogens (Howell 1989).

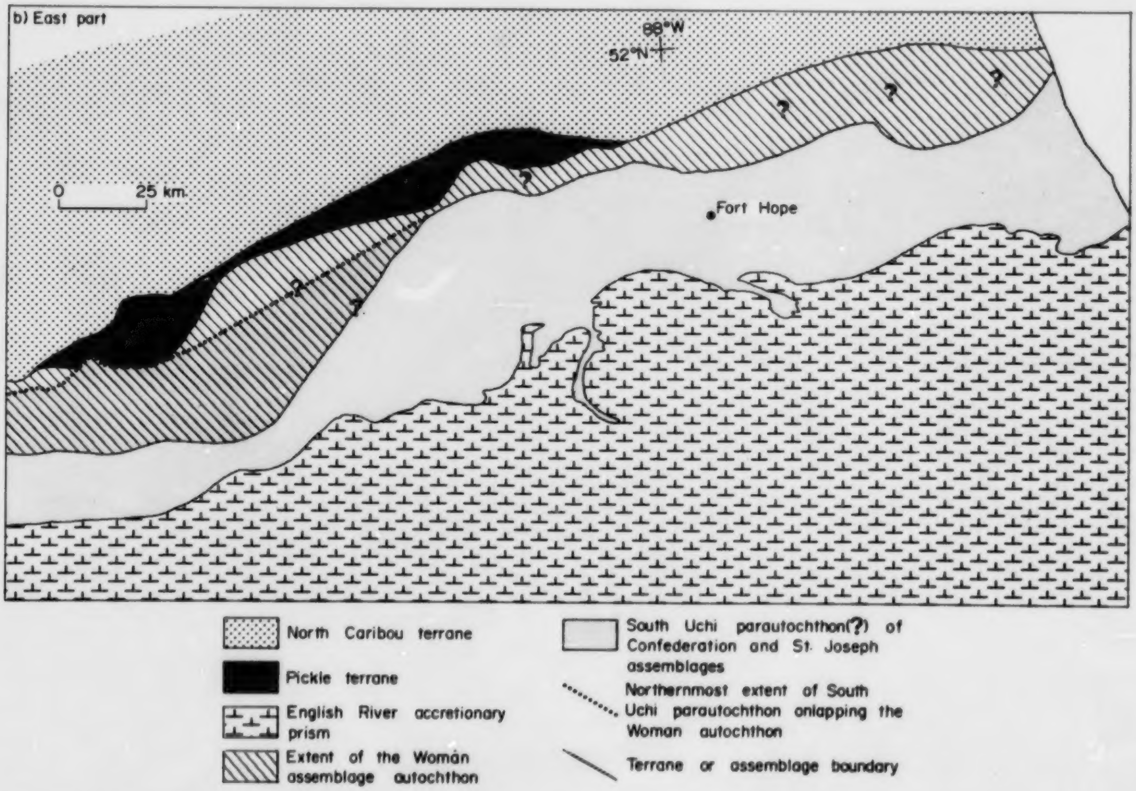
b) East part
88°W
52°N
?
?
?
?
?
?
0
25 km.
Fort Hope
North Caribou terrane
Pickle terrane
English River accretionary prism
Extent of the Woman assemblage autochthon
South Uchi parautochthon(?) of Confederation and St. Joseph assemblages
Northernmost extent of South Uchi parautochthon onlapping the Woman autochthon
Terrane or assemblage boundary

Figure 6.32. Schematic map of the main tectonic elements in the Uchi Subprovince. Viewed as composed of early accreted assemblages and subsequent arc and marginal basin additions along a continental margin, this subprovince contains the elements that formed the flank of an evolving microcontinent over a period of 300 million years. The initial stages involved formation of the North Caribou terrane, incorporating the pre–2.9 billion-year-old assemblages. The Pickle Crow and Bruce Channel assemblages are viewed as relicts of the Pickle terrane, accreted onto the North Caribou terrane prior to 2840 Ma. An autochthonous Woman assemblage is interpreted to have developed upon the margin of a North Caribou–Pickle composite terrane. Its original outboard extent is probably marked by the older 2890 million-year-old Pembina Tonalite Gneiss (*see* Figure 6.6). Later episodes of volcanism along this continental margin occurred to form the Confederation and St. Joseph assemblages before continental collison with a superterrane to the south. This collision caused northward displacement of these assemblages, which together form the South Uchi parautochthon, possibly interleaved locally with slices from the northern part of the English River assemblage.

WRENCH FAULTING

Wrench faulting characterizes the terminating stage of the Kenoran Orogeny as it affected the Uchi Subprovince. During this orogeny, shortening and thickening of the crust probably occurred largely as a consequence of convergence and collision of the superterranes directed along a northwestward principal axis, oblique to the westward-striking superterrane boundaries. The structural styles in different parts of the greenstone belts imply that this deformation was partitioned into crustal thickening (thrusting and folding) and wrenching components. Evidence in the northern part of the subprovince for the transpressive nature of the penetrative, ductile deformation is limited at present to areas, such as the Meen–Dempster greenstone belt, showing block rotation (Hale and Lloyd 1990) and associated orogen-subparallel stretching lineations (Stott and Wilson 1986b). But the principal evidence is based on the orientation of late-stage, regional-scale wrench (transcurrent) faults (e.g., Schwerdtner et al. 1979; Osmani and Stott 1988). Deformation near the Uchi–English River subprovince boundary was partitioned into: 1) shortening to produce tight folds with moderately east-plunging axes and stretching lineations in a narrow domain close to the boundary (Stott et al. 1987a), comparable to domains along the north margin of the Wabigoon Subprovince (*see* Blackburn et al., this volume) and the north margin of the Shebandowan greenstone belt, Wawa Subprovince (*see* Williams, Stott, Heather et al., this volume); and 2) transcurrent, simple shear in the Sydney Lake–Lake St. Joseph fault zone and in local, subparallel shear zones nearby within the greenstone belts (e.g., D_2 shear zones of the southern part of the Birch–Uchi greenstone belt). This transcurrent shear continued after the 2701 Ma crystallization of the Blackstone pluton. This late phase of deformation, concentrated in narrow zones of brittle-ductile shear, is also expressed in the 3 km wide Bear Head fault zone (*see* Figures 6.23a and 6.23b).

It is estimated that the main phase of Kenoran orogenesis in the Uchi Subprovince, the "Uchian phase" (Stott et al. 1989), occurred between 2713 and 2701 Ma. It predates the culmination of orogenesis in the southern part of the Superior Province, the "Shebandowanian phase" (Stott et al. 1989), which occurred at 2689 to 2684 Ma (Corfu and Stott 1986) in the Shebandowan greenstone belt and in the southern Wabigoon Subprovince (Davis et al. 1989; *see also* Williams, Stott, Heather et al., this volume). Most of the granitic plutons intruded after the culmination of the Uchian phase of the orogeny, from about 2704 to 2693 Ma. The Uchian phase is interpreted from features in the Lake St. Joseph area in the following paragraphs and schematically summarized in Figure 6.33.

1. The 2713 million-year-old St. Joseph volcanism marks the final phase of "southward" growth of a continental margin for the Uchi–Sachigo superterrane. Oblique "northward" subduction of an oceanic plate along this convergent margin is inferred. Convergence of another microcontinent (Winnipeg River and Wabigoon subprovinces) towards the Uchi–Sachigo superterrane ultimately produced a collision.
2. Much of the deformation in the southern half of the Uchi continental margin involved thrust interleaving of volcanic strata since 2722 Ma and during continental collision after 2713 Ma.
3. Exhumation and erosion of the converging microcontinents accompanied this deformation, forming the Eagle Island submarine fan prior to thrusting along the Uchi–English River boundary, which removed a substantial volume of St. Joseph assemblage rocks.
4. The culminating stage of collision produced folding along the leading edge of the Uchi–Sachigo superterrane, in the English River fore-arc accretionary prism, the Winnipeg River terrane (not shown), and along the northern margin of the Wabigoon composite terrane. Folds plunged steeply in the volcanic belts and moderately eastward in the metasedimentary prism and the gneissic Winnipeg River terrane. Ductile material in the English River prism was laterally displaced and deformed by a Wabigoon indentor of continental crust (in the Savant Lake region), producing the sinistral Miniss River wrench fault, coinciding with dextral wrench movement in the Sydney Lake–Lake St. Joseph fault zone. Wrench fault orientations and senses of displacement across the Superior Province correspond to northwest-directed transpression.
5. Collision and welding of microcontinents resulted in southward stepping of the subduction zone, rapidly followed by collision between Wabigoon and Wawa microcontinents during the Shebandowanian phase of the Kenoran Orogeny.

The Uchian and Shebandowanian orogenic phases, which could be viewed alternatively as separate micro-orogenies following rapidly one after another, culminated in major transpressive collision events, arising from oblique convergence of superterranes (Stott et al. 1987) and resulting in construction of the Superior Province. A synthesis of this construction, in the form of a cladogram, is presented in Williams, Stott, Thurston et al. (this volume), summarizing an interpretive sequence of tectonic events and the terrane-assemblage construction of the craton.

TECTONIC INTERPRETATION

An early history of accretionary growth in the northern part of the Uchi Subprovince is proposed to represent Stage 1 in a three-stage history of crustal growth and deformation that encompassed the whole subprovince. Stage 1 is followed by a better preserved record of crustal growth and deformation that dominates the southern half of the Uchi Subprovince, involving Stage 2—magmatism along a convergent continental margin, and Stage 3—microcontinent collision. Interpreted and speculative tectonic elements of the Uchi Subprovince arising from this three-stage history are given in Figures 6.32 and 6.34.

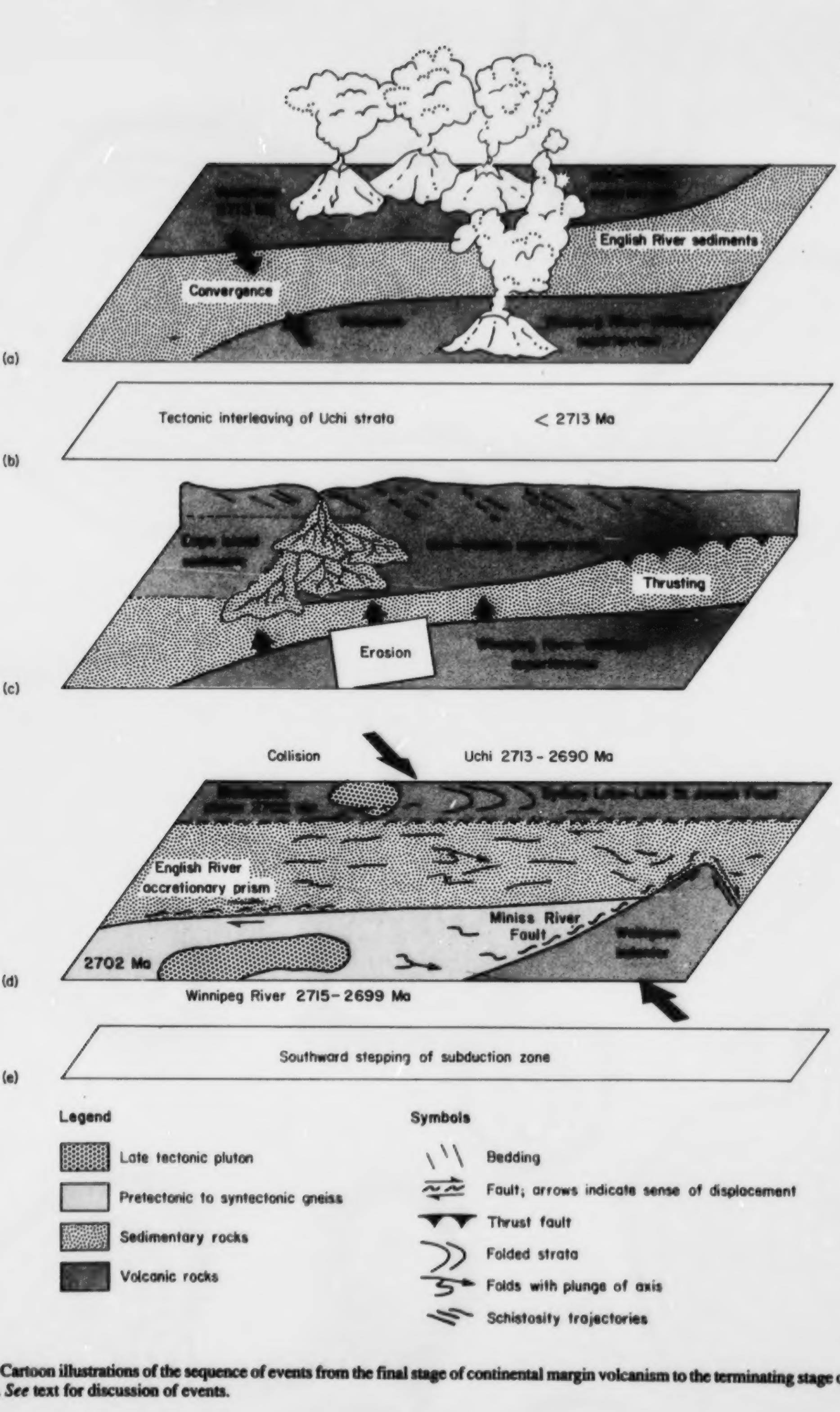
English River sediments
Convergence
(a)
Tectonic interleaving of Uchi strata
< 2713 Ma
(b)
Thrusting
Erosion
(c)
Collision
Uchi 2713 - 2690 Ma
English River accretionary prism
Miniss River Fault
2702 Ma
(d)
Winnipeg River 2715-2699 Ma
Southward stepping of subduction zone
(e)
Legend
Late tectonic pluton
Pretectonic to syntectonic gneiss
Sedimentary rocks
Volcanic rocks
Symbols
Bedding
Fault; arrows indicate sense of displacement
Thrust fault
Folded strata
Folds with plunge of axis
Schistosity trajectories

Figure 6.33. Cartoon illustrations of the sequence of events from the final stage of continental margin volcanism to the terminating stage of the Kenoran Orogeny. *See* text for discussion of events.

Stage 1. Accreting Terranes to form a Microcontinent

There is a growing enthusiasm amongst Archean researchers for postulating that accretionary tectonic processes, as observed in modern crustal settings, were also involved in producing the similar patterns of tectonic assemblages, structures, ages and sequences of events in Archean terranes. However, the evidence of growth of the entire Uchi Subprovince by accretion of terranes (in the sense of Howell 1989) is not presently compelling, owing in part to limited detailed studies of potential terrane boundaries in this region, but also to observations that are more consistent with

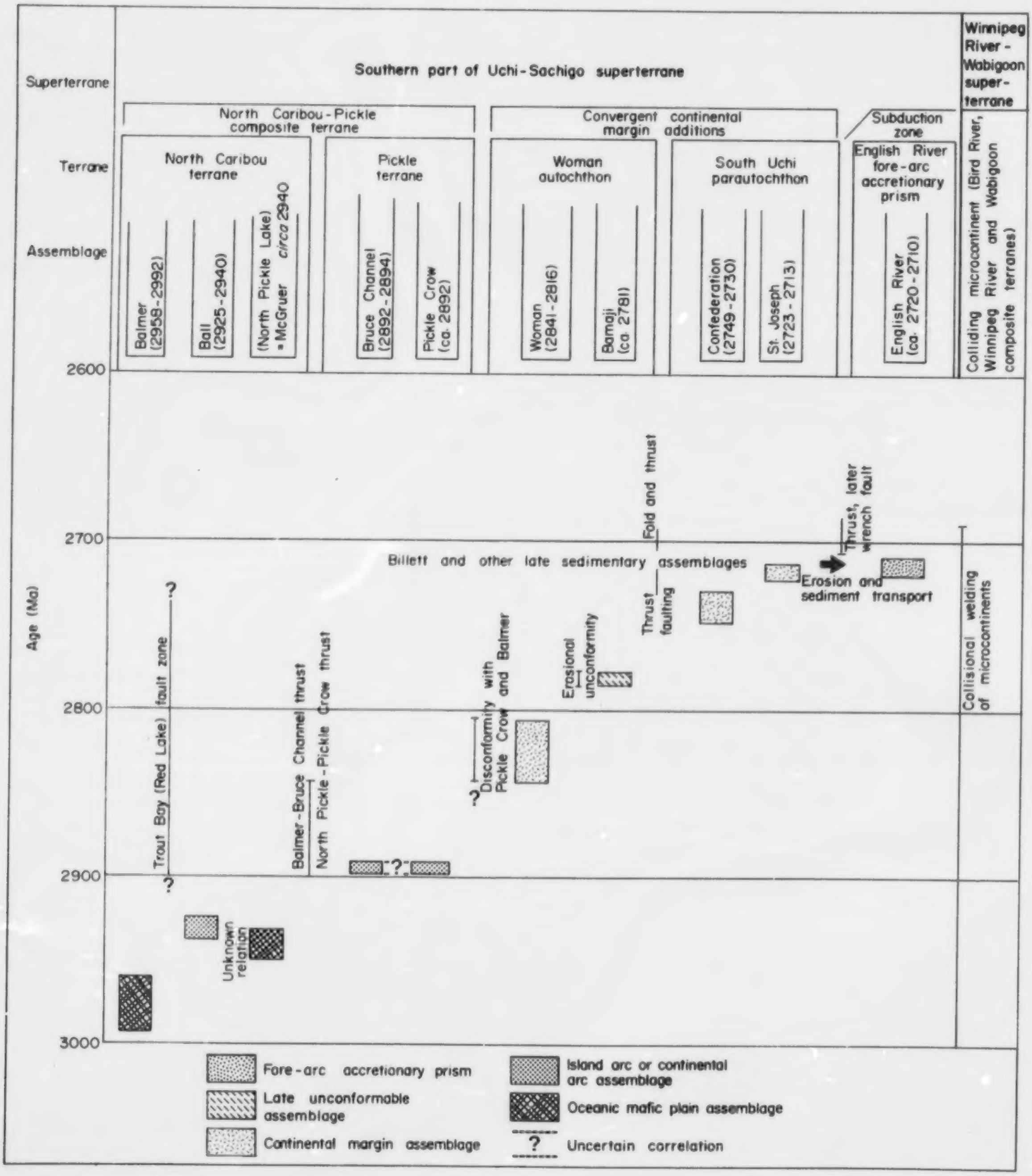

Figure 6.34. A time chart summarizing the main tectonic elements of Figure 6.32 as they were interpreted to have assembled in time.

another scenario. The authors, in agreement with Thurston, Osmani et al. (this volume), suggest that the northern parts of the subprovince, including the northern part of the Pickle Lake and Miminiska–Fort Hope greenstone belts, and much of the Red Lake greenstone belt and its slivers extending into the Berens River Subprovince (Stone 1988, 1990), are part of the North Caribou terrane (*see* Figure 6.32). Stratigraphic, aeromagnetic and age correlations between parts of the Uchi and Sachigo subprovinces support the concept of a large North Caribou terrane (*see* Figure 6.32). The region termed the Berens River Subprovince simply represents a deeper crustal level of this terrane and is dominated by late tectonic plutons, exhumed during the Uchian phase of the Kenoran Orogeny. This terrane is flanked on the south by the Pickle Crow assemblage and its apparent continuation into the Miminiska–Fort Hope greenstone belt. It is tectonically underlain by the Bruce Channel assemblage in the Red Lake greenstone belt. The structures in the Pickle Crow assemblage strike at an oblique angle to structures in the northern half of the Pickle Lake greenstone belt and similar discordance is evident in the northern Miminiska–Fort Hope greenstone belt. One could, therefore, postulate that the Pickle Crow assemblage and its synchronous counterpart, the Bruce Channel assemblage, in the Red Lake greenstone belt, are remnants of a 2890 million-year-old terrane (referred to here as the Pickle terrane) which accreted to the North Caribou terrane. The timing of such accretion is not clear, but it is limited to precede the disconformable juxtaposition of the 2840 million-year-old Woman assemblage against the Pickle Crow assemblage in the Pickle Lake greenstone belt and against the Balmer assemblage in the Birch–Uchi greenstone belt (*see* Figure 6.3). The development of the Woman assemblage represents the beginning of Stage 2.

Stage 2. A Convergent Continental Margin

It is apparent, in this scenario, that the Woman assemblage has an autochthonous origin. This assemblage is spatially associated with tonalite in the Trout Lake batholith, which has intruded rocks of the Balmer assemblage in the Birch–Uchi greenstone belt (based on relationships reported by Noble et al. 1985, p.76); as was argued in this chapter, this spatial relationship implies that the Woman assemblage more likely developed as an arc-marginal basin complex upon the edge of an older continent (a North Caribou–Pickle composite terrane?) and was not an intraoceanic terrane transported and accreted upon a conceivable North Caribou microcontinent. Consistent with this scenario is the presence of older, 2887 million-year-old tonalite gneiss, in the Lake St. Joseph batholith south of the Woman assemblage rocks, which probably formed part of the North Caribou–Pickle composite terrane but separated from the main mass of this terrane when the Woman arc complex formed. The Woman assemblage is proposed to have developed along the margin of the North Caribou–Pickle composite terrane as shown in Figure 6.32, although its present distribution may reflect some telescoping by compression during subsequent deformation events.

A similar argument can be made regarding the Confederation and St. Joseph assemblages. A volcanic vent in western Lake St. Joseph (Clifford and McNutt 1971) was episodically active around 2734 Ma and later at about 2723 Ma, forming the Confederation and St. Joseph assemblages. Plutonic bodies of similar age to these assemblages (e.g., Hammerton Lake pluton and Kawashe Lake stock) intruded older volcanic rocks in the northern part of the subprovince. The implication is that if the volcanism and plutonism are tectonically related, the Confederation and St. Joseph assemblages were formed from episodes of magmatism on the margin of a growing composite terrane that included the Woman assemblage of an earlier episode. The leading edge of this developing continent is inferred, from comparable features along the Quetico–Wabigoon subprovince boundary (*see* Blackburn et al., this volume), to include a fore-arc accretionary prism—the English River Subprovince—and to have overridden a "northward" (using present azimuth references) subducting oceanic plate until it ultimately collided with continental crust from the "south", the Winnipeg River and Wabigoon subprovinces.

An analogue is needed that accommodates the limitations to the original setting of Woman and younger assemblage volcanism, implied in the presence of synchronous plutonic phases intruding older crust. The setting of a Cretaceous island arc–marginal basin complex that developed along the western margin of southern South America partly satisfies the features observed in the Uchi Subprovince. This is the "Rocas Verdes" marginal basin complex of Chile that Tarney et al. (1976) advocated in very general terms as a model for the evolution of Archean greenstone belts. There are aspects of contrast, including the volume of andesitic volcanism in the island arc, which are not present in the Uchi assemblages. However, some elements of this continental marginal setting merit detailed comparison in future with the geological framework of the southern half of the Uchi Subprovince.

Stage 3. Continental Collision between Superterranes

The evidence for displacement and deformation, summarized in the previous section, especially for the southern half of the Uchi Subprovince, shows that the long, but episodic history of crustal growth along the Uchi–Sachigo convergent continental margin, including the late-stage history of the English River assemblage as a fore-arc accretionary prism, came to an end when this superterrane collided with a superterrane converging from the south (in the present-day azimuth)—probably consisting of the Winnipeg River and Wabigoon subprovince blocks. The styles of deformation during the Uchian collision include initial "thin-skin" thrusting, leading to "thick-skin" folding and thrusting that was accompanied by 1) downdip displacements near the leading edge of the Uchi–Sachigo continent, and 2) moderately plunging, orogen-parallel stretching lineations farther inboard. If the Confederation and St. Joseph assemblages developed as suggested, along the

Uchi margin of a microcontinent, their structural configurations including the northward-closing, lobate, east half of the Birch–Uchi greenstone belt, indicate that they formed what shall be called the South Uchi parautochthon (*see* Figures 6.32 and 6.34), with separate dismembered slices as small allochthons, as in the northern part of the Red Lake greenstone belt. The simplified picture that has so far emerged from the Uchi Subprovince indicates that the asymmetric distribution of volcanic assemblages and the styles and sequence of deformation especially at the terminating stage of collision, are similar to those documented in Phanerozoic orogens (e.g., Vauchez and Nicolas 1991).

Problems for Future Consideration

There is an abundance of opportunity for research in the Uchi Subprovince on all scales of investigation. Much of the Uchi story outlined in this chapter is based on inferences derived from stratigraphic and geochronological patterns and requires testing in several potentially fruitful areas. Areas that have sufficient bedrock exposure to permit detailed investigations include the Bee Lake belt and Red Lake greenstone belt, western Lake St. Joseph and parts of the Meen–Dempster greenstone belt. Of special importance are the thrust faults inferred from the age relationships of the stacked and uptilted assemblages; the crustal evolution of the Uchi Subprovince as a "fold and thrust belt" remains to be documented in detail. Detailed investigations are also warranted on the timing of gold mineralization in the Pickle Lake greenstone belt where evidence of gold mineralization that predates the Kenoran Orogeny is inviting. Not the least, the relationship between the tectonic settings and the igneous rock record remains to be told. Questions of regional-scale relationships to pursue include the following:

1. To what extent are the northeastward- and northwestward-trending greenstone belts, such as the Pickle Lake greenstone belt, part of an older fabric that evolved within the North Caribou terrane?
2. Do the northeast-trending long axes of granitoid batholiths, in the northern half of the subprovince, reflect an earlier (pre–Kenoran) orogenic flattening history or have the batholiths been rotated?
3. How far north can we trace a penetrative Kenoran orogenic shortening? Evidence in the Pickle Lake greenstone belt, for example, implies that older fabrics survive and that there may be an "orogenic front", north of which the expression of the Kenoran Orogeny is restricted to nonpenetrative fault dislocations and block rotations.
4. To what extent is the Woman assemblage a parautochthonous unit, which grew on the flanks of an evolving continent and was later partly dislocated and transported northward?
5. Can a better case be made for the dislocation and northward transportation of the Confederation assemblage? It has been suggested in this chapter that the eastern half of the Birch–Uchi greenstone belt is an allochthonous or parautochthonous "finger" protruding into the heart of the Uchi Subprovince; can this be confirmed? Are the greenstone belts and segments that are seemingly "out of place", such as the 2730 million-year-old volcanic sequence of the northern part of the Red Lake greenstone belt, or the 2749 million-year-old Lang Lake belt, allochthonous slices that were transported to the northern half of the subprovince?
6. What is the origin of the sedimentary assemblages in the midst of the greenstone belts? For example, is the Billett assemblage a foreland basin sequence, formed during *circa* 2709 Ma collision of the Uchi–Sachigo superterrane with a superterrane from the south?
7. To what extent is the formation of granite-greenstone subprovinces a product of terrane accretion, as recorded in the North American Cordillera, or a product of growth by continental arc magmatism, viewed by some workers as the principal process of growth in the Peruvian Andes (Cobbing *in* Pitcher et al. 1985)? In either case, there is clear evidence that collision of crustal blocks (superterranes), during the Kenoran Orogeny, sufficiently displaced and deformed crust as to make the resolution of this question more of a challenge.
8. Notwithstanding the position taken in this chapter, should the Kenoran Orogeny be viewed as composed of orogenic phases, the "Uchian" and "Shebandowanian" for example, or are each of these phases the process of separate orogens, rapidly assembled to form the Superior craton? Is our difficulty, in comparing the orogenic patterns of the Archean with those of subsequent eons, one of scale as well as rate of subduction and process?
9. It is apparent, from the crustal record and the inferred sequence of events, that the process of growth in each superterrane of northwestern Ontario terminated when the superterranes collided and welded together to form an Archean supercontinent (of which the Superior Province is a piece), effectively closing the Archean chapter of active plate convergence. Will the applications of high-precision U-Pb geochronology and stratigraphic analysis eventually permit the correlation of individual subprovinces amongst the various Archean cratons in the world? How long are they? There are sufficient contrasting features amongst the granite-greenstone subprovinces in the Superior Province to suggest that distinctive regional patterns of stratigraphy and structure across Archean cratons in different continents can eventually be matched up (cf., Hoffman 1991).

CONCLUSIONS

The Uchi Subprovince is a ribbon-like part of greenstone-granite crust composed of remnants that record protracted, episodic history of growth through Archean time from about 3 Ga to about 2.69 Ga. The most pronounced crust-forming periods, involving magmatism and sedimentation, occurred during 2992 to 2925 Ma, 2840 to 2806 Ma, and 2750 to 2713 Ma. Only the northernmost part of the subprovince, including part of the Red Lake greenstone belt and the Berens River batholithic belt, is interpreted as part of an older 2.9 to 3.0 billion-year-old crust in the North Caribou terrane. Possible remnants of a later terrane, the Pickle terrane, may have accreted to the proto-North Caribou continent before 2.84 Ga. However, subsequent volcanic assemblages and accompanying plutons most likely developed within a convergent continental margin. A crust-thickening period, involving thrust stacking and folding of volcanic and sedimentary assemblages, occurred mainly during the interval 2730 to 2700 Ma.

The post–2.84 Ga history of the Uchi Subprovince is best viewed as the record of an evolving Andean-type continental margin with a long history of crustal growth by continental arc magmatism. The record of clastic sedimentation associated with this magmatism is limited and reflects the products derived largely from erosion of volcanic edifices. This stage was terminated by continental collision with a superterrane to the south at about 2710 to 2700 Ma (the Uchian phase of the Kenoran Orogeny). This collision produced a "fold and thrust belt" encompassing most of the Uchi Subprovince. Northward-directed thrusting and stacking of volcanic strata were accompanied by folding and transcurrent faulting, and substantial exhumation and erosion of the continental crust to form synorogenic flysch-type basins; north-northwest-directed transpressive deformation was concentrated in the terminating stages of deformation along the leading edge of the continent, where back thrusting along a precursor of the Sydney Lake–Lake St. Joseph Fault changed to trancurrent shear. Accompanying and following the most penetrative phase of this collision event was a thermal event that produced a sizeable volume of plutonic magma to form numerous plutons within the subprovince and a plutonic belt analogous in scale to the Coastal Batholith of Peru. The longevity and significance of the Uchi Subprovince, as the leading edge of a continent that suffered the brunt of collision, is mirrored in the pronounced geophysical anomalies that characterize it and the underlying Archean mantle.

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Chapter 7

English River Subprovince

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Abstract

In Ontario, the English River Subprovince is 800 km long, 50 km wide, and forms a linear belt in the northwestern part of the Superior Province. The subprovince is composed predominantly of highly metamorphosed and migmatized clastic sedimentary rocks that are as young as 2698 Ma, minor metavolcanic rocks in the Melchett Lake metavolcanic-metasedimentary and Separation Lake greenstone belts and tonalitic to granitic intrusive rocks that range from 2.65 to 2.70 billion years in age. The supracrustal component mainly consists of metamorphosed wacke and mudstone. These rocks were originally deposited as marine turbidites in submarine fan and abyssal environments which formed a fore-arc basin along the southern margin of the Uchi Subprovince.

D_1 deformation resulted in the formation of recumbent thrusts and folds. D_2 deformation gave rise to a north-directed, layer-normal compression which caused dips of early planar fabrics to steepen to subvertical orientations. Abukuma-type regional metamorphism followed this deformation and developed maximum pressure-temperature conditions of 750°C and 0.54 GPa. This metamorphic event may have initiated several melt-producing episodes in the metasedimentary rocks between 2692 and 2660 Ma.

The basin in which rocks of the English River Subprovince occur may have developed as a fore-arc accretionary wedge. The thickness of the crust was increased by the D_1 imbricate stacking of units that comprised the sedimentary wedge, and the interleaving of this material with tonalite slabs greater than 3 billion years in age. Imbricate stacking of the sedimentary wedge and the interleaving of this material with tonalite slabs may have resulted from progressive convergence with a sialic microcontinent. This microcontinent is represented to the south of the English River Subprovince by the Winnipeg River Subprovince. Late Archean structural evolution in the English River Subprovince was marked by ductile to brittle deformation involving dextral, transcurrent motion along major subprovince boundary faults and along faults that transect the subprovince.

Mineral deposit types which occur in the area include: extensive iron formation of wacke turbidite association; rare-element pegmatites in low- and medium-grade subprovince boundary zones; radioelement-enriched pegmatites situated in medium- and high-grade areas in the western English River Subprovince; copper-nickel-cobalt-palladium-platinum-bearing sulphides in meta-ultramafic pods in the Werner Lake fault zone; zinc-lead-copper-silver volcanogenic massive sulphide mineralization restricted to the Melchett Lake metavolcanic-metasedimentary belt and the Separation Lake–Lac Seul area; and rare gold mineralization.

INTRODUCTION

Geologic studies have been carried out in the English River Subprovince for more than a century (Figure 7.1). These studies commenced in 1872 with the work of the Geological Survey of Canada (Bell 1873; Dowling 1896) and continued with that of the Ontario Geological Survey (Prest 1944; Pye 1968). It remained for Wilson and Brisbin (1963), however, to propose a subdivision of the crust in the northwestern Superior Province in which the distinctiveness of the high-grade metasedimentary, migmatite-dominant subprovince described in this chapter was recognized. Wilson and Brisbin (1963) called this subdivision the "English River Gneiss Belt". Additional geologic and geographic definitions by Stockwell (1964, 1970) and Wilson (1971), respectively, coined the names English River Subprovince and English River Block. The high metamorphic grade, abundance of paragneisses, subordinate metavolcanic rocks and presence of regionally extensive boundary faults were among the criteria used by these workers to distinguish the 800 km long English River Subprovince from adjacent, lower grade, greenstone-rich subprovinces. The recognition of 2 belts, having distinctively different lithic characters within the then English River Subprovince, was a further significant development. Regional mapping by Beakhouse (1974a, 1977), Breaks, Bond, Harris et al. (1975) and Breaks et al. (1978) provided a basis for dividing the English River Subprovince into a northern metasedimentary, migmatite-dominated belt, named the Ear Falls–Manigotagan gneiss belt or northern supracrustal domain, and a southern pluton-rich terrain that featured a paucity of supracrustal rocks, called the Winnipeg River Batholithic Belt or southern plutonic domain. Geophysical observations led to a proposal by Hall and Brisbin (1982) for separate subprovince status for each lithologic division. Card and Ciesielski (1986) proposed a nomenclature that retained the name English River Subprovince, representing the northern supracrustal domain, which is used in this review. The southern plutonic domain became known as the Winnipeg River Subprovince and is described by Beakhouse (this volume).

This chapter provides a description and analysis of the geologic, structural, metamorphic and mineral deposit features of the English River Subprovince. The chapter proposes an evolutionary model accounting for the geologic

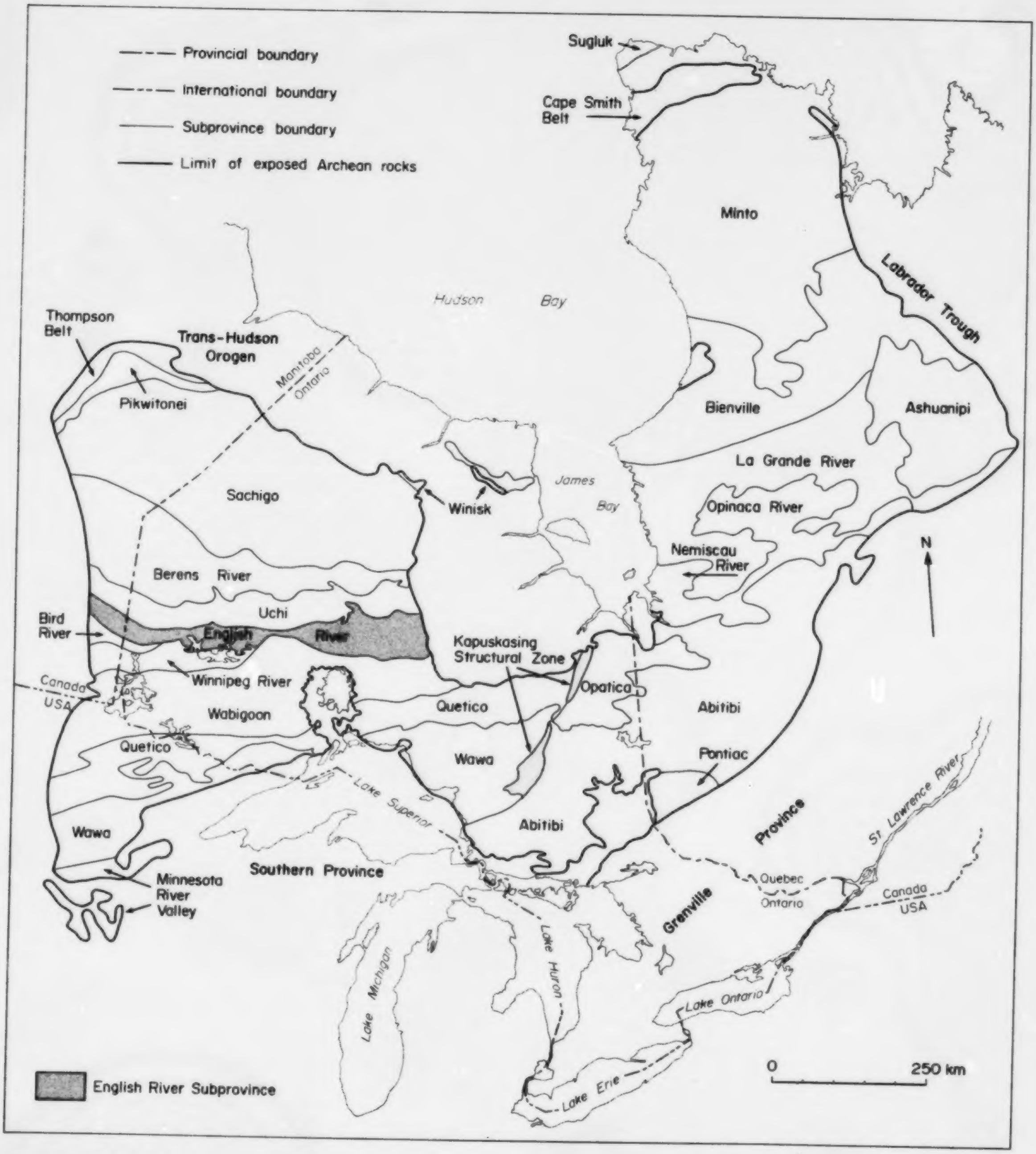

Provincial boundary
International boundary
Subprovince boundary
Limit of exposed Archean rocks
Sugluk
Cape Smith Belt
Minto
Labrador Trough
Hudson Bay
Thompson Belt
Trans-Hudson Orogen
Manitoba
Ontario
Pikwitonei
Bienville
Ashuanipi
La Grande River
Sachigo
James Bay
Winisk
Opinaca River
Nemiscau River
N
Berens River
Uchi
Bird River
English River
Kapuskasing Structural Zone
Winnipeg River
Canada
USA
Opatica
Wabigoon
Quetico
Abitibi
Quetico
Wawa
Pontiac
Lake Superior
Province
Abitibi
St. Lawrence River
Wawa
Southern Province
Minnesota River Valley
Quebec
Ontario
Canada
USA
Grenville
Lake Huron
Lake Michigan
Lake Ontario
Lake Erie
0
250 km
English River Subprovince

Figure 7.1. Location of the English River Subprovince in relation to neighbouring subprovinces of the Superior Province (*modified from* Card and Ciesielski 1986).

observations. All unreferenced U-Pb zircon ages in this review were determined by Fernando Corfu, of the Jack Satterly Geochronological Laboratory, Royal Ontario Museum, Toronto.

Geologic Overview

The English River Subprovince is an elongate, east-striking, high-grade, clastic metasedimentary-plutonic belt similar to the Quetico Subprovince in the Superior Province (*see* Williams, this volume). The subprovince can be traced for 800 km from the eastern shore of Lake Winnipeg, where McRitchie and Weber (1971) described the correlative Manigotagan gneisses eastward to long. 86°W, where it is unconformably overlain by Upper Ordovician sedimentary rocks of the James Bay Lowlands. Aeromagnetic data (*see* Maps 2585 and 2586, map case) show that anomalies characteristic of the English River Subprovince continue another 175 km east, to the inferred extension of the northeast-striking Gravel River fault (Williams 1989), beneath the Phanerozoic cover of the James Bay Lowlands. Card (1990) suggested that the Nemiscau River and Opinica River subprovinces (*see* Figure 7.1) may represent the eastward extension of the English River Subprovince metasedimentary mass into Quebec. The width of the English River Subprovince is quite variable, and ranges from 1.5 to 51 km and averages about 30 km. A pronounced narrowing is evident in the vicinity of Pashkokogan Lake, which likely relates to displacement along intersecting northeast- and northwest-striking faults.

The English River Subprovince has been interpreted as an interarc sedimentary basin rimmed by quasi-contemporaneous volcanic rocks (Dwibedi 1968; Beakhouse 1974a, 1974b, 1977; Ayres 1978; Breaks et al. 1978) and more recently, in a plate tectonic context, has been interpreted as an accretionary prism (Hoffman 1989). Interbedded wacke and pelite, and their migmatitic derivatives, compose 60% of the area of the subprovince. The maximum age of sedimentation is poorly constrained, but it continued until 2698 Ma, which is the age of the youngest detrital zircon, and by analogy, the youngest possible age for tonalite intrusion into the metasedimentary rocks.

Metavolcanic rocks compose only about 2% of the English River Subprovince by area. The largest metavolcanic belt in the English River Subprovince occurs at Melchett Lake (Figure 7.2). This belt differs lithologically from greenstone belts of adjacent greenstone-dominated subprovinces, in that 80% of its metavolcanic rocks are felsic to intermediate in composition (Bond 1979; Bond and Foster 1981a, 1981b).

Numerous intrabelt batholiths, stocks and allied dikes were emplaced between 2698 and 2560±40 Ma (Krogh et al. 1976; Wooden and Goodwin 1980; Corfu, Royal Ontario Museum, personal communication, 1989). These plutonic rocks account for most of the remainder of the English River Subprovince and are divisible into 5 distinct groups in terms of their relative and absolute ages, and are listed below in order of decreasing abundance.

1. gneissic tonalite suite (in part, as old as 3170$^{+20}_{-5}$ Ma)
2. tonalite-trondhjemite-granodiorite suite (2665±20 to greater than 3000 Ma)
3. peraluminous granite-granodiorite suite (2668 to 2692 Ma)
4. biotite granite-granodiorite suite (2560±40 to 2698 Ma)
5. mafic-ultramafic plutonic suite (absolute age unknown)

The boundary zone relationships of the English River Subprovince with adjacent subprovinces are variable in character. These relationships can be tectonically conformable, intrusive and can be severed by layer-parallel or obliquely striking regional faults. In some cases, these relationships are contentious and not fully determined. The northern boundary of the English River Subprovince is exclusively defined by its contact with the Uchi Subprovince (*see* Figure 7.2). Original stratigraphic relationships at the English River Subprovince–Uchi Subprovince interface are largely obscured by the Sydney Lake–Lake St. Joseph Fault (*see* Figure 7.2). Where uninterrupted lithofacies sequences across the boundary are seen, English River Subprovince clastic metasedimentary rocks are interpreted to lie on (Breaks et al. 1978; Nunes and Thurston 1980; Wallace 1981a, 1981b) or beneath (Jensen 1989, 1991a, 1991b) 2714 to 2738 million-year-old metavolcanic units of the Uchi Subprovince. The southern boundary is largely defined by the northern edges of greenstone belts in the Wabigoon Subprovince, for example, east of Smoothrock Lake (*see* Figure 7.2) and the Bird River Subprovince west of Helder Lake (*see* Figure 7.2). Between these 2 areas, in the Chamberlain Narrows region, the English River Subprovince lies in juxtaposition with 2690±15 to 3170$^{+20}_{-5}$ million-year-old granitoid rocks of the Winnipeg River Subprovince (Beakhouse 1983; Corfu 1988; Sanborn-Barrie 1988). This poorly constrained contact of the English River Subprovince with the Winnipeg River Subprovince is represented by a combination of curvilinear, conformable contacts with foliated and gneissic tonalitic and sedimentary rocks of similar ages and structural style, which have undergone a complex deformation history (Breaks et al. 1978; Sanborn-Barrie 1988). The interdigitating contacts are caused by a combination of deformation and late tectonic to posttectonic emplacement of massive granodiorite to granite suite plutons. East of long. 93°30′W, relationships at the southern boundary of the English River Subprovince with the Winnipeg River Subprovince, become vague (*see* Figure 7.2) because of an absence of boundary defining greenstone belts of the Wabigoon Subprovince between Savant Lake and Lonebreast Bay in Smoothrock Lake (Sage et al. 1974). Granulite-grade mineral assemblages in mafic volcanic rocks, intrusive rocks and clastic metasedimentary rocks and the presence of widespread orthopyroxene-bearing tonalites coincide with a high-contrast linear magnetic signature (*see* maps 2585 and 2586, map case), and may be used as an indication that the Winnipeg River Subprovince probably continues eastwards to the Whitewater Lake area. Granulite-grade rocks are not typical of the Wabigoon Subprovince, so this area of tonalitic rocks in the south of the English River Subprovince could either be part of the

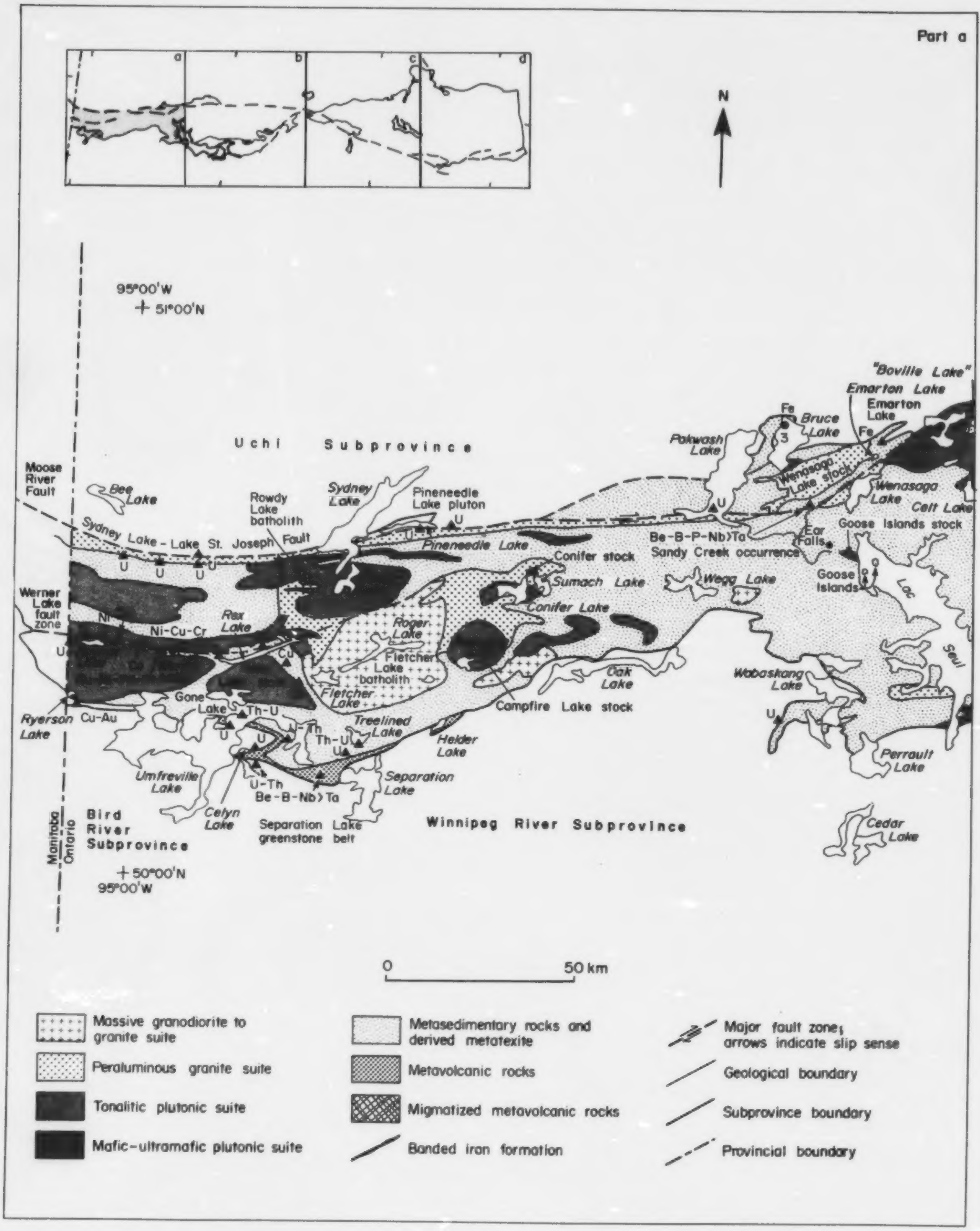
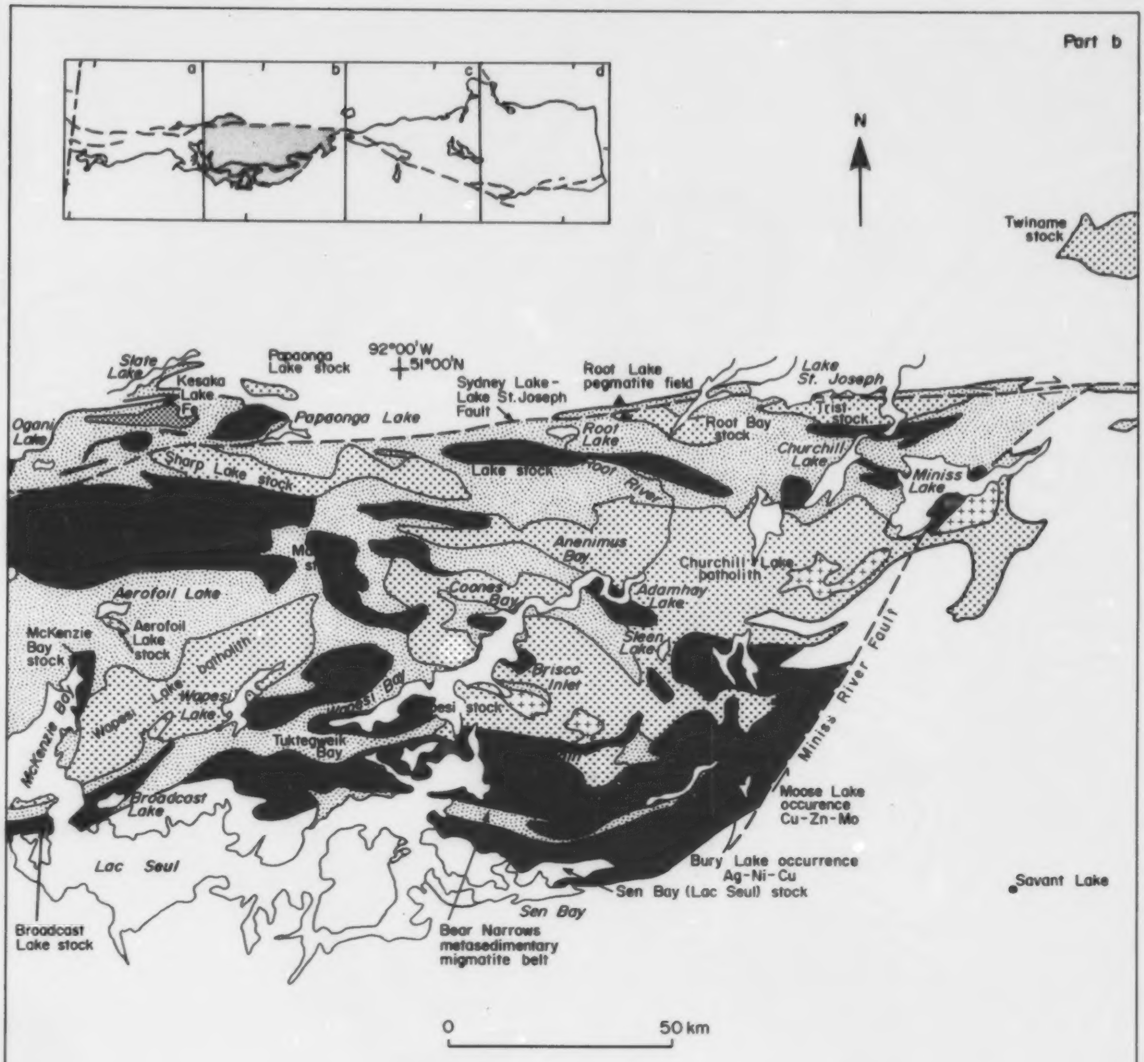

Part a
a
b
c
d
N
95°00'W
51°00'N
Uchi Subprovince
Moose River Fault
Bee Lake
Sydney Lake
Rowdy Lake batholith
Pineneedle Lake pluton
Sydney Lake - Lake St. Joseph Fault
Pineneedle Lake
Conifer stock
Sumach Lake
Conifer Lake
Werner Lake fault zone
Ni
Ni-Cu-Cr
Rex Lake
Roger Lake
Fletcher Lake batholith
Fletcher Lake
Cu
Gone Lake
Th-U
Ryerson Lake
Cu-Au
Treelined Lake
Th-U
Helder Lake
Campfire Lake stock
Oak Lake
Umfreville Lake
Celyn Lake
U-Th
Be-B-Nb>Ta
Separation Lake
Separation Lake greenstone belt
Bird River Subprovince
Manitoba
Ontario
Winnipeg River Subprovince
50°00'N
95°00'W
"Boville Lake"
Emarton Lake
Emarton Lake
Fe
Bruce Lake
Pakwash Lake
3
Wenasaga Lake stock
Wenasaga Lake
Celt Lake
Goose Islands stock
Be-B-P-Nb>Ta
Sandy Creek occurrence
Ear Falls
Wegg Lake
Goose Islands
Lac Seul
Wabaskang Lake
Perrault Lake
Cedar Lake
U
0
50 km
Massive granodiorite to granite suite
Peraluminous granite suite
Tonalitic plutonic suite
Mafic-ultramafic plutonic suite
Metasedimentary rocks and derived metatexite
Metavolcanic rocks
Migmatized metavolcanic rocks
Banded iron formation
Major fault zone; arrows indicate slip sense
Geological boundary
Subprovince boundary
Provincial boundary

Figure 7.2a. General geology of the English River Subprovince showing place names referred to in the text, mineral occurrences and mines.


Part b
a
b
c
d
N
Twiname stock
Slate Lake
Kesaka Lake
Fe
Papaonga Lake stock
92°00'W
51°00'N
Sydney Lake - Lake St.Joseph Fault
Root Lake pegmatite field
Lake St. Joseph
Trist stock
Oganj Lake
Papaonga Lake
Root Lake
Root Bay stock
Churchill Lake
Miniss Lake
Sharp Lake stock
Lake stock
Root River
Anenimus Bay
Churchill Lake batholith
Aerofoil Lake
Coones Bay
Adamhay Lake
McKenzie Bay stock
Aerofoil Lake stock
Sleen Lake
Wapesi Lake batholith
Brisco Inlet
Wapesi Lake
Wapesi Bay
Wapesi stock
McKenzie Bay
Tukteawik Bay
Miniss River Fault
Moose Lake occurence Cu-Zn-Mo
Broadcast Lake
Bury Lake occurrence Ag-Ni-Cu
Lac Seul
Sen Bay (Lac Seul) stock
Savant Lake
Sen Bay
Broadcast Lake stock
Bear Narrows metasedimentary migmatite belt
0
50 km

▲Au Mineral occurrence

Past-producing mines:

1 Werner Lake Mine (1929-1932; 1940-1944)

2 Gordon Lake Mine (1962-1972)

3 Griffith Mine (1966-1979)

Abbreviations:

Ag	silver	Nb	niobium
Au	gold	Ni	nickel
B	boron	P	phosphorous
Be	beryllium	Pb	lead
Co	cobalt	Pd	palladium
Cr	chromium	Pt	platinum
Cs	cesium	Rb	rubidium
Cu	copper	Sn	tin
F	fluorine	Ta	tantalum
Fe	iron	Th	thorium
Li	lithium	U	uranium
Mo	molybdenum	W	tungsten
		Zn	zinc

Figure 7.2b. General geology of the English River Subprovince showing place names referred to in the text, mineral occurrences and mines.

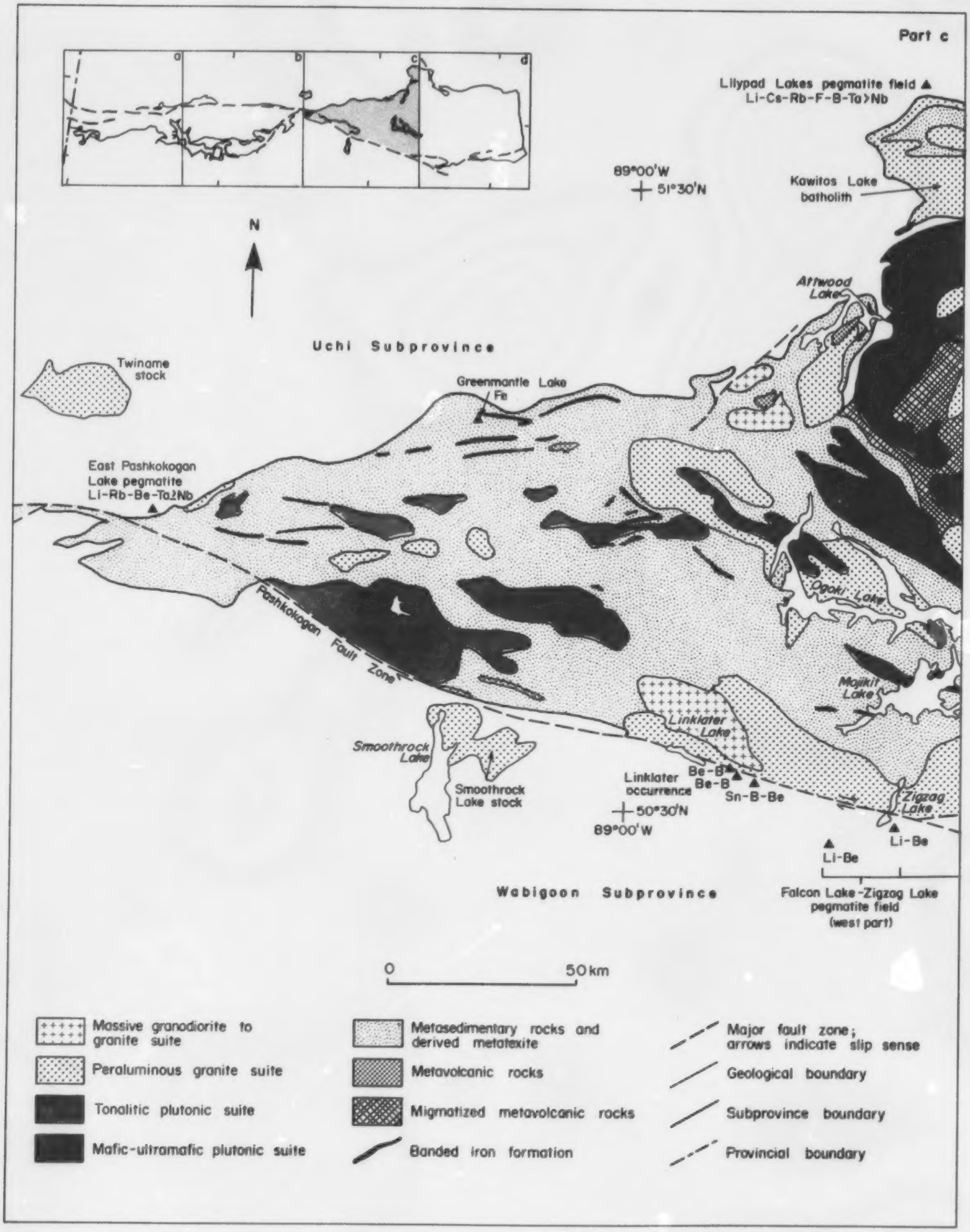
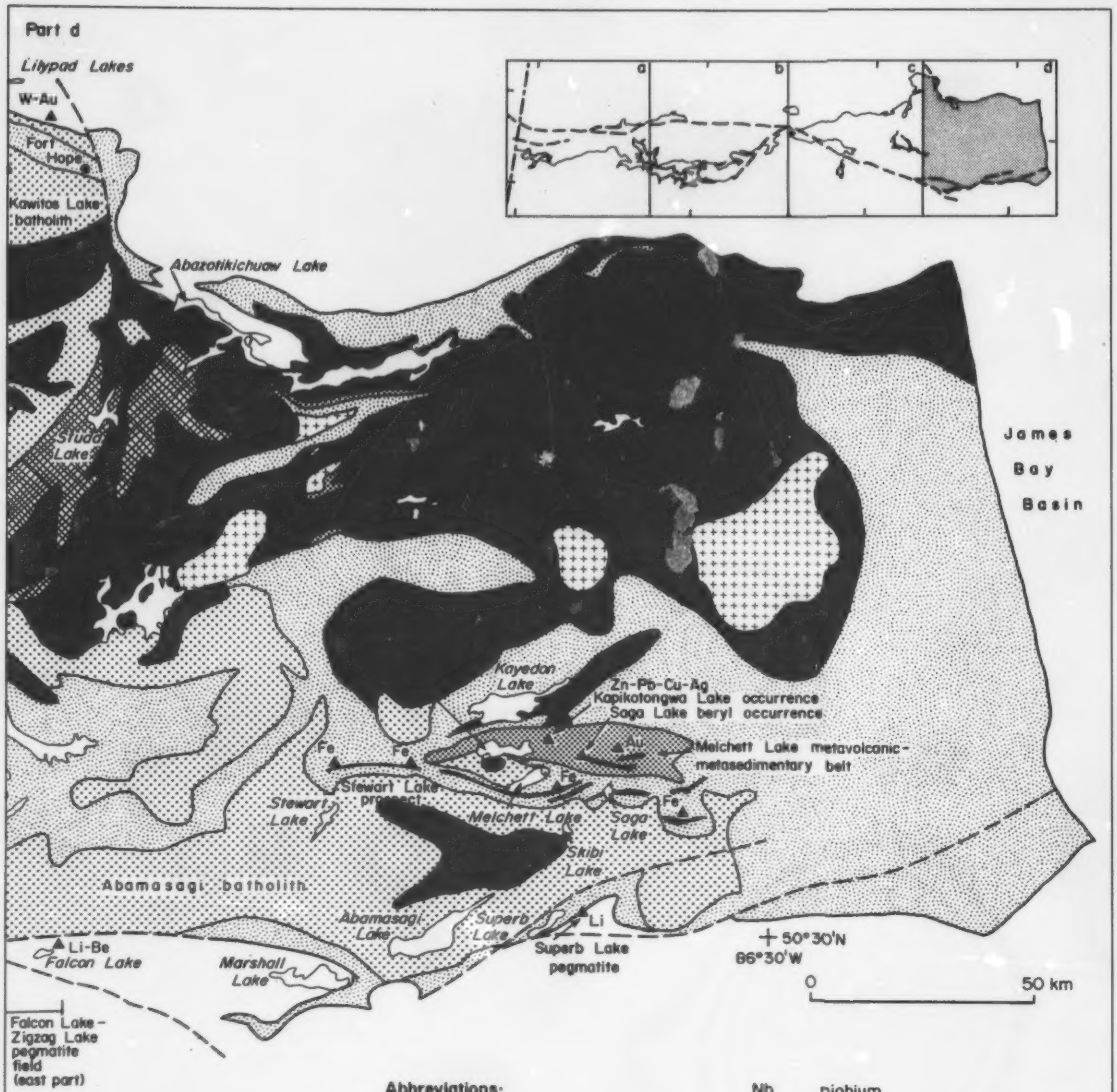

Part c
a
b
c
d
Lilypad Lakes pegmatite field ▲
Li-Cs-Rb-F-B-Ta>Nb
89°00'W
51°30'N
Kawitos Lake batholith
N
Attwood Lake
Uchi Subprovince
Twiname stock
Greenmantle Lake
Fe
East Pashkokogan Lake pegmatite
Li-Rb-Be-Ta≥Nb
Ogoki Lake
Pashkokogan Fault Zone
Majikit Lake
Linklater Lake
Smoothrock Lake
Smoothrock Lake stock
Linklater occurrence
Be-B
Be-B
Sn-B-Be
50°30'N
89°00'W
Zigzag Lake
Li-Be
Li-Be
Wabigoon Subprovince
Falcon Lake-Zigzag Lake pegmatite field (west part)
0
50 km
Massive granodiorite to granite suite
Peraluminous granite suite
Tonalitic plutonic suite
Mafic-ultramafic plutonic suite
Metasedimentary rocks and derived metatexite
Metavolcanic rocks
Migmatized metavolcanic rocks
Banded iron formation
Major fault zone; arrows indicate slip sense
Geological boundary
Subprovince boundary
Provincial boundary

Figure 7.2c. General geology of the English River Subprovince showing place names referred to in the text, mineral occurrences and mines.


Part d
Lilypad Lakes
W-Au
Fort Hope
Kawitos Lake batholith
Abazotikichuaw Lake
Studd Lake
a
b
c
d
James Bay Basin
Kayedan Lake
Zn-Pb-Cu-Ag
Kapikotongwa Lake occurrence
Saga Lake beryl occurrence
Au
Melchett Lake metavolcanic-metasedimentary belt
Fe
Fe
Fe
Fe
Stewart Lake prospect
Stewart Lake
Melchett Lake
Saga Lake
Skibi Lake
Abamasagi batholith
Abamasagi Lake
Superb Lake
Li
Superb Lake pegmatite
50°30'N
86°30'W
0
50 km
Li-Be
Falcon Lake
Marshall Lake
Falcon Lake - Zigzag Lake pegmatite field (east part)

▲Au Mineral occurrence

Past-producing mines:

1 Werner Lake Mine (1929-1932; 1940-1944)

2 Gordon Lake Mine (1962-1972)

3 Griffith Mine (1966-1979)

Abbreviations:

Ag	silver	Nb	niobium
Au	gold	Ni	nickel
B	boron	P	phosphorous
Be	beryllium	Pb	lead
Co	cobalt	Pd	palladium
Cr	chromium	Pt	platinum
Cs	cesium	Rb	rubidium
Cu	copper	Sn	tin
F	fluorine	Ta	tantalum
Fe	iron	Th	thorium
Li	lithium	U	uranium
Mo	molybdenum	W	tungsten
		Zn	zinc

Figure 7.2d. General geology of the English River Subprovince showing place names referred to in the text, mineral occurrences and mines.

Winnipeg River Subprovince or a pluton-dominated segment of the English River Subprovince.

Deformation of the English River Subprovince developed during the Kenoran Orogeny at *circa* 2.7 Ga and involved 3 periods of folding and a protracted period of brittle-ductile phenomena involving kinking, shear zone development, fracturing and cataclasis. The intensity of deformation diminished with time (McRitchie and Weber 1971) and became less penetrative.

The regional metamorphism which developed in the English River Subprovince during the Kenoran Orogeny is of the low-pressure, high-temperature Abukuma-type (Miyashiro 1961), and has a steep, (estimated) vertical metamorphic gradient of 35° to 50°C/km (Breaks et al. 1978). Where uninterrupted by boundary faults, metamorphic grade increases horizontally southwards from the English River Subprovince–Uchi Subprovince boundary and ranges from low-grade up to granulite-grade metamorphism. Boundary fault zones typically coincide with abrupt increases in grade (Figure 7.3). The English River Subprovince has undergone significant erosion and uplift, relative to the adjacent greenstone belts in the Uchi Subprovince.

Mineralization in the English River Subprovince exhibits chalcophile, siderophile and lithophile affinities. Mineralization is found in: 1) two types of granitic pegmatites enriched either in uranium and thorium or rare-elements (e.g., lithium, cesium, rubidium, beryllium and tantalum); 2) iron formation; 3) zinc-lead-copper volcanogenic massive sulphide; and 4) platinum-palladium-copper-nickel-cobalt mineralized mafic-ultramafic rocks; rare gold deposits also occur.

LITHOLOGIC DESCRIPTIONS

Field relationships and petrographic descriptions of English River Subprovince rock units are presented in this section, and are based on Breaks et al. (1978) and Breaks and Bond (in prep.).

CLASTIC METASEDIMENTARY ROCKS AND DERIVED MIGMATITES

Five lithofacies types have been regionally recognized within the most extensive map unit of the English River Subprovince. Two of these types predominate, are commonly interbedded and consist of: quartzofeldspathic rocks, considered to be metamorphosed wackes; and pelitic rocks, which are rich in micas and/or other peraluminous minerals, and are interpreted to have a mudstone parentage. Conglomerate, calc-silicate units and quartz-plagioclase arkose units contrast with this rather uniform sedimentary association. Banded iron formation of wacke-magnetite turbidite association, although volumetrically minor, occurs in laterally extensive units which are as much as 35 km along strike, and tend to be concentrated near the English River Subprovince–Uchi Subprovince interface (Breaks et al. 1978; Meyn and Palonen 1980; Sage and Breaks 1982). Thin, mafic-rich units, concordant with enclosing sedimentary rock, also occur sporadically. These units may represent recrystallized metamorphosed mafic sheets, distally deposited basaltic tuff, or rocks of unknown protolith that were subjected to seafloor alteration.

Clastic Metasedimentary Rocks

Wacke is distinguished by being generally equigranular, massive to weakly foliated in appearance (Figures 7.4a and 7.4c) and having an essential mineralogy of quartz, plagioclase and biotite (Figure 7.5). Metamorphic grain size is fine to medium grained, commonly coarsening slightly in higher grade metamorphic zones. Porphyroblastic variants, marked by low modal percent of coarse-grained garnet or rarer cordierite, occur sporadically in medium- and high-grade zones. Relative to mudstone, wacke is richer in quartz and total feldspar, resulting in a lower colour index.

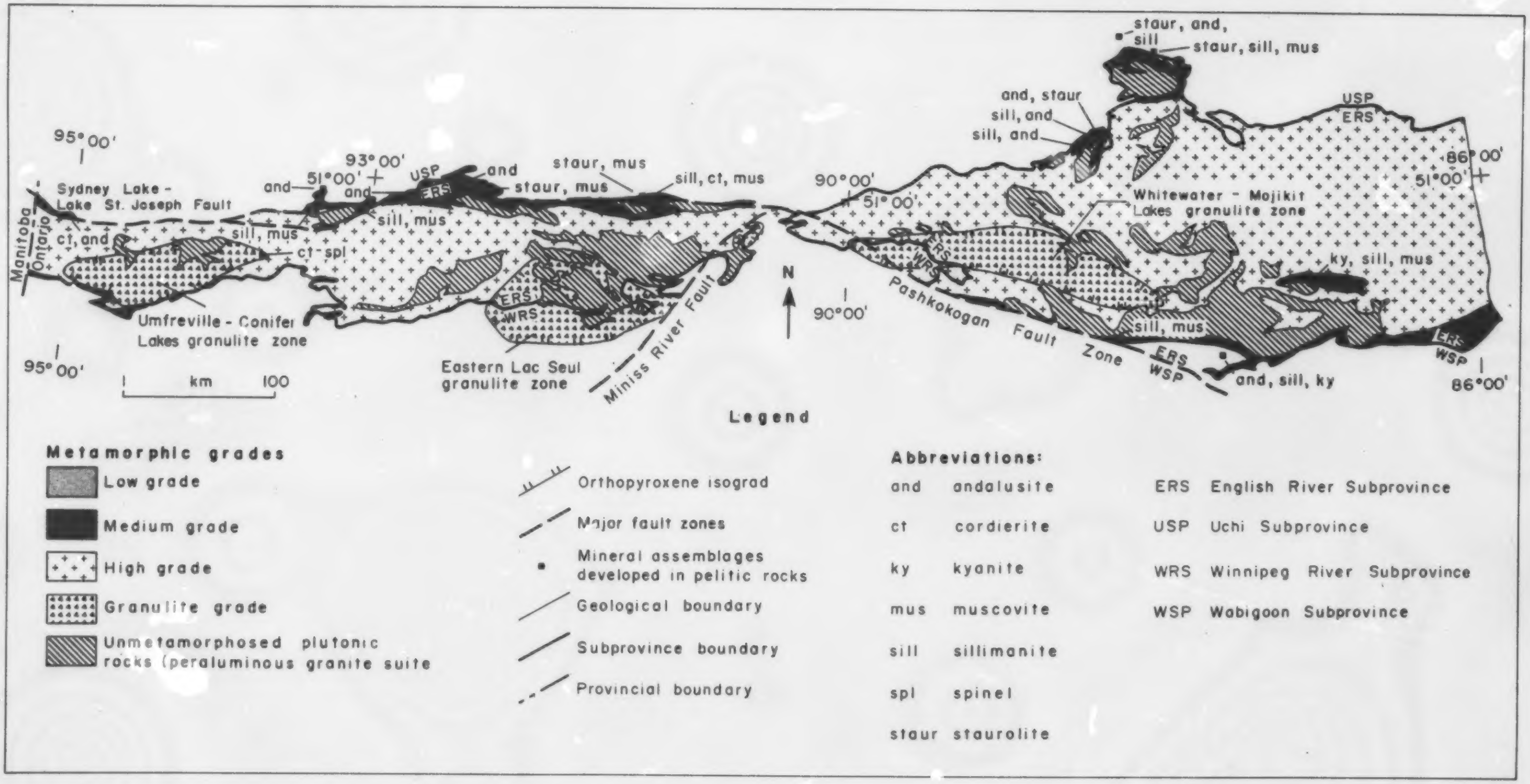
The mudstone is fine grained and equigranular in the low- and medium-grade zones. With increasing metamorphic grade, mudstone is rapidly converted into schistose, inequigranular, generally coarse-grained, porphyroblastic varieties (*see* Figure 7.4b). Metamorphism produced a regional distribution of essential mineral constituents: quartz, plagioclase, microcline, biotite, muscovite, chlorite, staurolite, andalusite, sillimanite, garnet, cordierite and orthopyroxene.

Arkose units are volumetrically insignificant, but nevertheless, these units are ubiquitous. These thinly bedded units are conspicuously leucocratic and exhibit a lighter weathered surface than the intercalated wackes or pelites.

Scarce conglomerate is dominantly found within the boundary zones with adjacent to greenstone-dominated subprovinces. Conglomerate occurrences are documented by Burwash (1923), Derry (1931), Breaks et al. (1978), van de Kamp and Beakhouse (1979), Meyn and Palonen (1980), Trueman (1980), Wallace (1981a, 1981b), Sutcliffe (1988) and Sanborn-Barrie (1988). Two conglomerate units are found at the contacts between metasedimentary migmatite and tonalite-trondhjemite-granodiorite suite rocks along the English River Subprovince–Winnipeg River Subprovince boundary in the Oak and Perrault lakes areas (Westerman 1977; Breaks et al. 1978; Sanborn-Barrie 1988).

Calc-silicate rocks represent the rarest lithofacies type; they are found only in the eastern Lac Seul area (*see* Figure 7.2). In this area, grey-green, diopside-rich units are locally associated with mafic metavolcanic rocks and chert-magnetite banded iron formation, or as andesine-diopside-magnetite layers in migmatized metasedimentary rocks (Harris and Goodwin 1976).

Primary sedimentary textures and structures, such as bedding in wacke and mudstone, are best preserved in boundary zone localities where fault-induced deformation is minimal (e.g., the western Lake St. Joseph and Papaonga Lake–Slate Lake areas; *see* Figure 7.2). Even within high- and granulite-grade metamorphic zones (*see* Figure 7.3), where migmatization is variable, bedding contacts between wacke and pelite are readily discernible, and are commonly enhanced by metamorphic recrystallization. Bedding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Metamorphic grades
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Medium grade
High grade
Granulite grade
Unmetamorphosed plutonic
rocks (peraluminous granite suite
Orthopyroxene isograd
Major fault zones
Mineral assemblages
developed in pelitic rocks
Geological boundary
Subprovince boundary
Provincial boundary
Abbreviations:
and andalusite
ct cordierite
ky kyanite
mus muscovite
sill sillimanite
spl spinel
staur staurolite
ERS English River Subprovince
USP Uchi Subprovince
WRS Winnipeg River Subprovince
WSP Wabigoon Subprovince

Figure 7.3. Generalized regional metamorphic zones of the English River Subprovince and neighbouring parts of the Winnipeg River Subprovince.

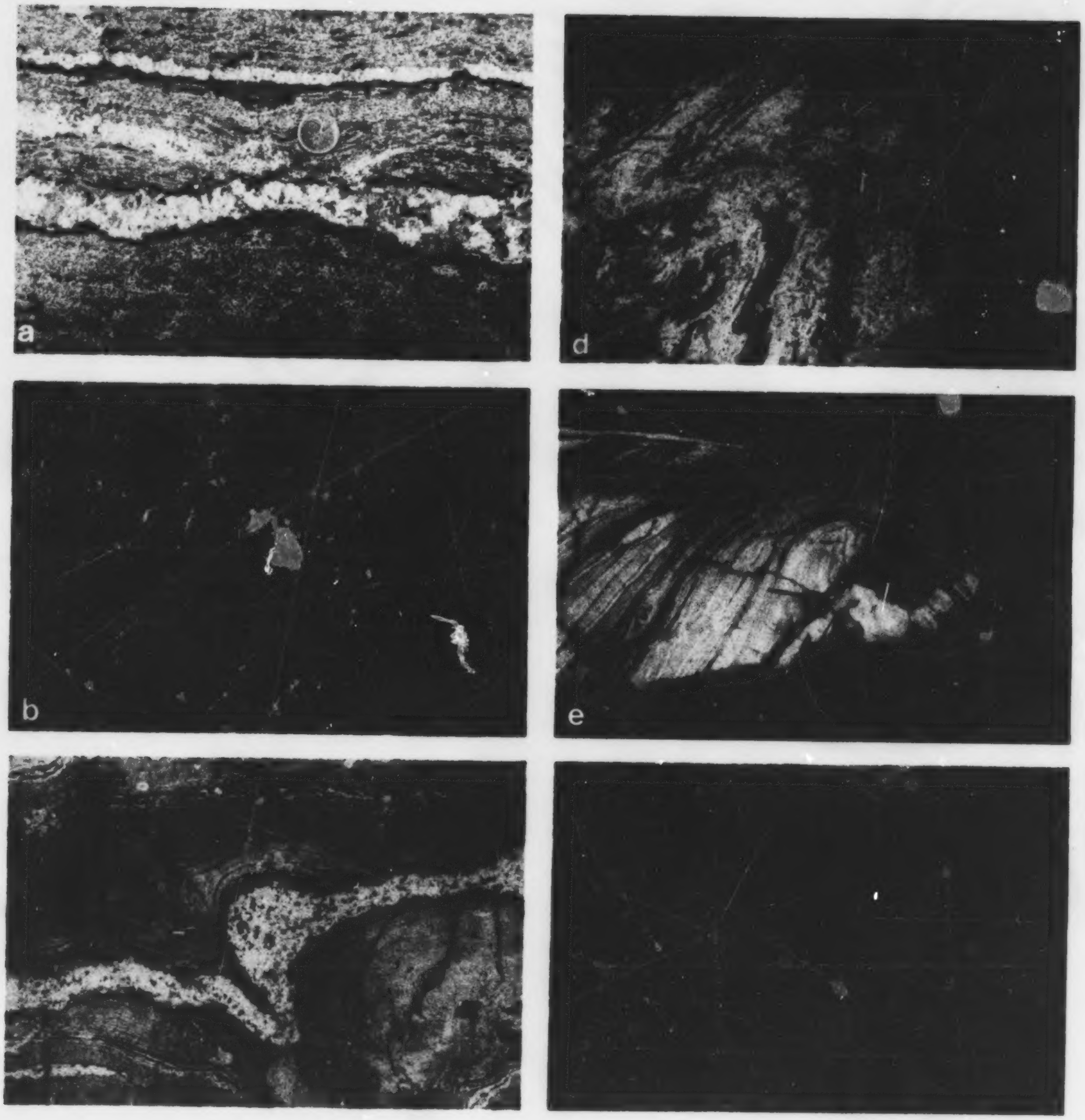

Figure 7.4. **a)** Metatexite with exceptionally thick biotite-rich melanosome selvages developed along cordierite leucosome + garnet + sillimanite + cordierite + potassium feldspar + biotite mesosome; Ontario Ministry of Natural Resources facilities on Lac Seul at Ear Falls. Coin is 2.5 cm in diameter; **b)** pelite layer in protometatexite containing abundant porphyroblasts of cordierite and garnet and podiform leucosome; Wegg Lake. Penknife is 10 cm long; **c)** apatite-biotite quartz syenite leucosome in metatexite, note development of biotite-rich melanosomes along leucosome contacts; Ontario Hydro dam site at Ear Falls. Penknife is 10 cm long; **d)** transition at Treelined Lake area from metatexite on right into inhomogeneous diatexite containing twisted rafts of metawacke and mesosome in cordierite-garnet-biotite leucosome. Hammer is 35 cm long; **e)** highly deformed tonalitic gneisses, arrow marks location of a narrow biotite-filled shear which has abruptly truncated the gneissosity; Tuktegweik Bay, Lac Seul. Outcrop is approximately 2 m high; and **f)** tonalitic gneiss from northwestern part of the Bluffy Lake batholith near "Boville Lake". Coin in the left centre of the photograph is 2.5 cm in diameter.

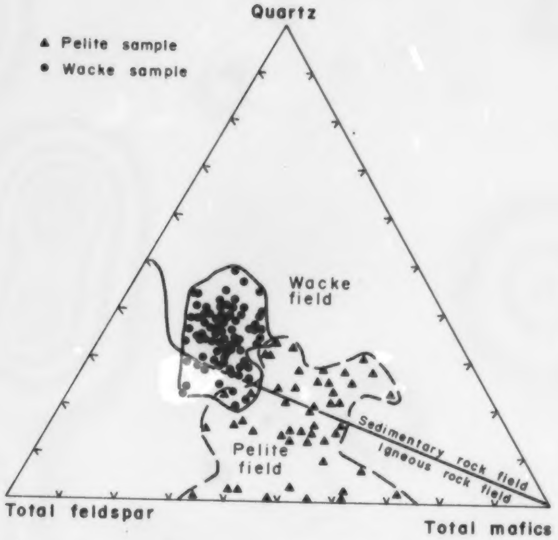

Figure 7.5. Modal variation of quartz, total feldspar and total mafic minerals for metasedimentary rocks of the English River Subprovince. Sedimentary and igneous rock compositional fields *modified from* van de Kamp and Beakhouse (1979).

thickness is variable, ranging from less than 1 cm to more than 2 m (van de Kamp and Beakhouse 1979), but generally bedding is in the 5 cm to 1.5 m range. Less widely preserved primary sedimentary structures include internal laminations (Breaks et al. 1978; Meyn and Palonen 1980), cross-stratification (Kindle 1932; Breaks, Bond, Harris et al. 1975; van de Kamp and Beakhouse 1979), clastic dykes (Meyn and Palonen 1980; Sage and Breaks 1982), load casts (Meyn and Palonen 1980; Card 1983), slump balls, rip-ups and ripples (Meyn and Palonen 1980; Card 1983) and flame structures (Card 1983). Textural features such as graded bedding are generally not well preserved within the high- and granulite-grade zones. In rare cases, graded bedding can be deduced from the preferential concentration of aluminous metamorphic phases (garnet, biotite, sillimanite and cordierite) towards one part of a bed (e.g., van deKamp and Beakhouse 1979).

ENVIRONMENT OF SEDIMENTARY DEPOSITION

The depositional environment of the English River Subprovince metasedimentary mass is difficult to decipher due to the ravages of metamorphism, migmatization and deformation. Van de Kamp and Beakhouse (1979) inferred a turbidite origin for the migmatized metasedimentary rocks, and that these rocks were possibly formed in a submarine fan environment. The regional dominance of metamorphosed equivalents of wacke and mudstone possibly reflects derivation from former Bouma A–E turbidite couplets; fining- and thinning-upwards sequences of wacke, siltstone and shale, containing sporadic conglomerate and iron formation are well preserved in the English River Subprovince–Uchi Subprovince boundary zone (e.g., Meyn and Palonen 1980; Wallace 1981a, 1981b), and correspond with the resedimented turbidite facies of Ojakangas (1985). Meyn and Palonen (1980) considered that these rocks developed as a southward-prograding, lower submarine fan (Walker 1978), implying a source terrane to the north.

Migmatized Metasedimentary Rocks and Derived Granites

High-grade metamorphism, imposed upon the clastic metasedimentary rocks between 2692 and 2660 Ma, produced a diversity of migmatite types. These types, along with terminology of migmatite components in these rocks (Table 7.1) conform with the descriptive, nongenetic usage currently in use (Mehnert 1973; Ashworth 1985). A field classification scheme (e.g., Mehnert 1973; Brown 1973) was developed for the metasedimentary migmatites (*see* Table 7.1). It consists of several gradational stages marked by the amount of leucosome increasing in relation to mesosome. This classification has been useful in the reconnaissance mapping of variably migmatized sedimentary belts such as that in the English River Subprovince and has furthered the recognition of the peraluminous, one- or two-mica granite suite, whose origin is closely linked with such belts. Note that on the Bedrock Geology of Ontario (*see* Maps 2542 and 2543, map case) the term "muscovite-bearing granitic rocks" includes peraluminous granitic rocks in the English River Subprovince which usually do not contain muscovite. Several episodes of granitic leucosome generation both *in situ* and of magmatic injection origin are recognized (Breaks and Bond, in prep.) in relation to the deformation events D_1, D_2, D_3 and D_4. Therefore, a given exposure of metasedimentary migmatite may be complex and not always amenable to classification, due to the abundance of overprinted leucosomes and intensity of associated deformation events.

METATEXITE

The most common migmatite type is metatexite; it is encountered in the southward transition from the low- and medium-grade English River Subprovince–Uchi Subprovince boundary zone metasedimentary rocks towards those of the high-grade zone (*see* Figures 7.3 and 7.4a). Protometatexite, as defined by Breaks et al. (1978), is a variant of metatexite and exhibits an incipient segregation of leucosome and is typified by less than 10% leucosome patches (*see* Figure 7.4b). These are specifically confined to pelitic mesosomes and are generally absent from intercalated wacke mesosomes. The segregations have not been incorporated into the foliation-controlled, continuous leucosome layers which are normally indicative of metatexite.

Development of *in situ* syenitic leucosomes occurs on a limited scale (*see* Figure 7.4c) and may reflect derivation of an anatectic melt from a metasedimentary protolith depleted in quartz but enriched in muscovite and potassium feldspar (subordinate albitic plagioclase may also be present). The apatite content of 5 to 10% in the Ear Falls leucosome, abnormally high relative to typical English River Subprovince granite leucosome, may point to the effect of P_2O_5 on

Table 7.1. Classification and descriptive features of migmatitic metasedimentary and related granitic rocks in the English River Subprovince (migmatite terminology *modified from* Brown 1973; Mehnert 1973; Breaks et al. 1978; Ashworth 1985).

Migmatitic Type	Leucosome Content of Migmatite or Granite	Field Relations, Petrography and Migmatite Structure	Possible Genetic Processes
Protometatexite	<10%	Characterized by low quantities of podiform and lensoid granite, leucosome mainly confined to high-grade, porphyroblastic pelite mesosome. Intercalated metawacke typically absent of leucosome.	Incipient anatexis selectively developed in low calcium, potassium feldspar + albite + quartz pelitic protoliths. Solid-state segregation in chemically or tectonically induced gradients (Ramberg 1949, 1952; Barth 1952; Reitan 1956; Kretz 1967; Sawyer and Robin 1986) and metasomatism (Misch 1968; Olsen 1985) could also be important.
Metatexite	10 to 60%	Migmatites of stromatic and/or phlebitic structure produced by striped leucosomes controlled by foliation and relict bedding in mesosome host; white leucosomes generally hololeucocratic, coarse grained, mainly granite (s.s.) and less commonly granodiorite, trondhjemite and quartz syenite. Melanosomes commonly present, 1 to 5 mm thick, rich in biotite with lesser cordierite, sillimanite and plagioclase.	Leucosomes may be derived by *in situ* anatexis (Holmquist 1921; Winkler and von Platen 1958) or by subsolidus mechanisms (*see* references in protometatexite category). Injection of mobile granitic magma along a planar fabric also important (Sederholm 1907).
Inhomogeneous Diatexite	60 to 90%	Granitic rocks featured by a conspicuous abundance of inclusions of metatexite and unmigmatized metawacke, and schlieren rich in biotite (possible melanosome). Nonmetasedimentary inclusions such as metaultramafic rock are rare. Schollen structure is especially characteristic. Leucosome typically coarse grained, white, hypidiomorphic granular, massive to moderately foliated. Colour index commonly 10 to 30.	Anatexis is dominant; increase in granitic melt component *vis–a–vis* mesosome component led to mechanical disruption of the stromatic leucosome–melanosome–mesosome framework of metatexite.
Homogeneous Diatexite	>90%	Granitic rock which is typically massive, homogeneous, and seriate, hypidiomorphic to granular, coarse grained to pegmatitic and colour index typically less than 10. Wide range in modal composition; granite (s.s.) is most common. Rare inclusions of metatexite and unmigmatized metawacke mesosome.	Extensive anatexis and probable amalgamation of granitic melt into magmatic masses which were capable of ascent into the crust above the zone of fusion of the metasedimentary protolith.

the derivation of syenitic melt compositions (Wyllie and Tuttle 1964).

DIATEXITE

Increased amounts of leucosome (*see* Table 7.1) progressively disaggregate the layered migmatite framework of metatexite, and transforming the rock into "schollen" and "nebulitic" diatexitic migmatite structures (Mehnert 1973). The proportion of leucosome in the transitional inhomogeneous diatexite stage, ranges from 40 to 60% (*see* Figure 7.4d), and may have been derived by *in situ* anatexis and/or by injection along the planar mesosome fabric.

Diatexite occurs on a regional scale (*see* high-grade zones on Figure 7.3) with a total area covering about 5000 km² in the English River Subprovince. These rocks are particularly abundant within, or proximal to, some granulite-grade metamorphic zones (*see* Figure 7.3). An advanced stage of fusion during the breakdown of biotite in the protolith metasedimentary rocks may have played an important role in producing the characteristic rock textures in diatexite (Vielzeuf and Holloway 1988).

Recently, diatexites in the English River Subprovince have been reclassified into the peraluminous granite suite (Breaks 1989, p.28–31), in view of their distinctive chemistry and accessory mineralogy, and their strong spatial relationship to metasedimentary-rich subprovinces. These rocks will be described more fully in the descriptive section of the granitic rock suites.

METAVOLCANIC ROCKS

Metavolcanic rocks are scarce in the English River Subprovince and are mainly confined to the Melchett Lake area (*see* Figure 7.2). These rocks also occur as thin units of amphibolite composition, intercalated with metasedimentary rocks, within the English River Subprovince–Winnipeg River Subprovince and English River Subprovince–Uchi Subprovince boundary zones. The constituent metavolcanic units of the 30 km long and 5 km wide Melchett Lake metavolcanic-metasedimentary belt include felsic and intermediate, dominantly subaqueous, pyroclastic flows, which compose about 90% of the volcanic assemblage and minor pillowed, massive, and brecciated mafic flows compose the remainder (Bond 1979; Bond and Foster 1981a, 1981b). Associated metasedimentary rock types include staurolite-, cordierite-, sillimanite- and muscovite-bearing pelite and metawacke (Bond and Foster 1981a, 1981b; Card 1983). The data on younging (Bond and Foster 1981a, 1981b) indicate that the supracrustal sequence faces south with clastic metasedimentary rocks and interbedded oxide facies banded iron formation lying at the top of the sequence.

Massive to gneissic amphibolite interpreted to be largely volcanogenic, occurs in many localities scattered along a 250 km strike length of the English River Subprovince–Winnipeg River Subprovince boundary east of Separation Lake (*see* Figure 7.2). These occurrences were interpreted by Breaks et al. (1978) to represent remnants of a once continuous volcanic belt, possibly of a minimum age of 2745±5 Ma (Timmins et al. 1985). This age is interpreted from a tentative correlation of the Lamprey Falls Formation with the Separation Lake greenstone belt. One of the best exposed localities is in eastern Lac Seul where numerous amphibolite units up to 0.5 by 6 km are dispersed throughout the Chamberlain Narrows batholith (*see* Figure 7.2; Breaks and Bond, in prep.).

INTRUSIVE ROCKS

Plutonic intrusive rocks within the English River Subprovince can be divided into 4 felsic to intermediate suites, and 1 mafic to ultramafic suite. This division is based upon field relations and petrographic and chemical characteristics (Breaks et al. 1978; Breaks and Bond, in prep.) and is supported by a few age determinations (Table 7.2). These rocks account for about 40% of the English River Subprovince, and vary in size from small dikes and sills, to batholiths as large as 2200 km² in area. Less than 1% of the English River Subprovince is underlain by mafic to ultramafic intrusive rocks.

Gneissic Tonalite Suite

These complex, highly deformed, metaplutonic rocks form only a small proportion of the English River Subprovince relative to the Winnipeg River Subprovince, where they account for 20% of the area (Breaks et al. 1978; Beakhouse, this volume). Formerly called the "gneissic granitoid suite" by Breaks et al. (1978), the current name more accurately reflects the dominance of tonalite in the compositional spectrum (Figures 7.6a, 7.7a; *see* Table 7.2). The age of this suite within the English River Subprovince is undetermined, but the nearest dated compositional equivalent occurs in the Cedar Lake area of the Winnipeg River Subprovince, 15 km south of the English River Subprovince, where Corfu (1988) obtained an age of 3170^{+20}_{-5} Ma. Evidence has not been forthcoming to indicate how much of the tonalite gneiss in the English River Subprovince is this age.

The gneissic tonalite suite (*see* Figures 7.6a and 7.7a) comprises a heterogeneous array of intricately, interleaved tonalitic units which vary in colour index, and contain derived migmatitic leucosomes, minor supracrustal enclaves and dike-derived amphibolite. The suite is invaded by several generations of dikes of the massive granodiorite-granite and tonalite-trondhjemite-granodiorite suites. Tonalitic gneisses are characterized by penetrative planar or folded compositional banding (*see* Figure 7.4e).

The gneissic tonalite suite is contained within the complex tonalite intrusions of the tonalite-trondhjemite-granodiorite suite that occurs within the interior of the English River Subprovince or, along segments of the English River Subprovince–Winnipeg River Subprovince boundary. An example of the former is the Bluffy Lake batholith (*see* Figures 7.2 and 7.4f) in which gneissic tonalite zones could either represent enclaves of older sialic crust, or consanguineous but more highly deformed parts of such plutons. Tonalitic gneisses associated with the subprovince boundary zones form shallow-dipping, straight gneiss zones along the contacts with thin supracrustal units.

Table 7.2. Field, petrographic, metallogenic and geochronologic features of intrusive suites in the English River Subprovince.

Intrusive Suite	Form and Distribution	Composition
Gneissic Tonalite	Small zones; <10 km^2 in area; in TTG* suite plutons and also within ERS–WRS* boundary zone.	Trondhjemite, tonalite, granodiorite; minor diorite and quartz diorite.
Trondhjemite-Tonalite-Granodiorite	Batholiths, stocks, and small dikes regionally distributed within metasedimentary migmatite and or along ERS–WRS* boundary zone.	Trondhjemite and tonalite predominate; minor granodiorite, diorite and quartz diorite; simple and complex plutons can be distinguished on basis of number of internal units (up to 10).
Massive Granodiorite to Granite	Mainly small stocks and dikes; batholiths are rare.	Dominantly granite and granodiorite; lesser alkali feldspar granite and quartz monzonite.
Peraluminous Granite	Batholiths and stocks of regional extent and numerous chemically fractionated pegmatitic dikes intrusive into subprovince boundary zone, low- and medium-grade supracrustal rocks.	Granite predominant; subordinate granodiorite, trondhjemite, quartz syenite and alkali feldspar granite
Mafic-Ultramafic	Tectonic inclusions, dikes and small to large stocks.	Diorite-quartz diorite-minor gabbro; gabbro-pyroxenite-peridotite; amphibolite.

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* *Abbreviations: TTG—trondhjemite-tonalite-granodiorite; ERS—English River Subprovince; WRS—Winnipeg River Subprovince; USP—Uchi Subprovince.*

Foliated Tonalite Suite (or Tonalite-Trondhjemite-Granodiorite Suite)

Rocks dominantly composed of tonalite and trondhjemite represent the most abundant igneous plutonic suite in the English River Subprovince and have furnished isotopic ages from 2.65 to 3.0 Ga (*see* Maps 2579 and 2580, map case). Minor associated rock units include granodiorite, diorite, and quartz diorite. Cumulatively, these rocks comprise about 25% of the English River Subprovince and are distributed uniformly (*see* Figure 7.2 and Table 7.2). There are 2 principal modes of occurrence: one mode is discrete, ovoid to elongate stocks and batholiths entirely enclosed by metasedimentary migmatite and the other mode forms more extensive, linear belts of batholithic dimensions, disposed along and near the English River Subprovince–Winnipeg River Subprovince boundary.

Relative to the chemically similar gneissic tonalite suite, the tonalite-trondhjemite-granodiorite suite is less compositionally and structurally complex and is relatively unmigmatized (*see* Figure 7.6b). Compositional banding is generally absent but these rocks exhibit a weak to strong metamorphic foliation and/or mineral lineation imparted by biotite, elongated quartz and, in mafic units, by alignment of hornblende. Some units may appear massive even in very old intrusive complexes, such as the greater than 3.0 billion-year-old Chamberlain Narrows batholith (*see* Figure 7.2). Breaks and Bond (in prep.) recognized 2 petrographic types of tonalite-trondhjemite-granodiorite intrusions: simple, compositionally uniform plutons; and complex, multiunit heterolithic masses.

Compositionally uniform plutons are mainly stocks composed of a few internal phases. The plutons strongly deflect banding in the host metasedimentary migmatites into concordance with their external plutonic contact. Examples include the Campfire Lake, Gone Lake (*see* Figure 7.2) and Adamhay Lake stocks (*see* Figure 7.9).

Complex, heterolithic masses of tonalite-trondhjemite-granodiorite are more abundant than the uniform type

Table 7.2. Field, petrographic, metallogenic and geochronologic features of intrusive suites in the English River Subprovince.

Petrographic Features	Age	Associated Mineralization	Examples
Banded (10 to 50 cm thick), highly deformed, grey to white, heterogeneous (up to 20 lithic units at outcrop scale); individual units are strongly foliated; several generations of migmatitic leucosome are typical; commonly fine to medium grained; recrystallized textures dominant.	Unresolved; possibly similar to 3170 million-year-old Cedar Lake Gneiss (Corfu 1988).	None known.	Narrow <1 km zones in Chamberlain Narrows batholith; northeast part of Bluffy Lake batholith.
Foliated and/or lineated to massive, grey to white, commonly medium grained, hypidiomorphic and allotriomorphic granular; locally potassium feldspar megacrystic; local patches of migmatitic leucosome may be present.	2.65 to >3.008 Ma (Wooden, NASA, unpublished data; Krogh et al. 1976).	None known	Chamberlain Narrows (Lac Seul), Bluffy Lake and Rowdy Lake batholiths; Otatakan Lake, Campfire Lake and Whitewater Lake stocks.
Mainly massive, locally foliated, pink, medium to coarse grained, equigranular, allotriomorphic to hypidiomorphic granular; pegmatite grain sizes are typical of the posttectonic dikes; potassium feldspar porphyritic texture found only in Fletcher Lake batholith; colour index commonly <10.	2698 to 2560 Ma (Corfu, Royal Ontario Museum, unpublished data; Krogh et al. 1976).	U-Th-enriched pegmatites.	Aerofoil Lake stock, Fletcher Lake batholith.
Massive to weakly foliated, locally cataclastic, white, medium grained to pegmatitic, normally hypidiomorphic granular and seriate, colour index <1 to 30.	2692 to 2668 Ma (Corfu, Royal Ontario Museum, unpublished data).	Rare–element pegmatites of Be-P-B, and Li-Rb-Be-Ta and Li-Cs-Rb–Be-Ta and Sn-Be specialization, derived from "fertile" peraluminous granites in ERS–USP[*] and ERS–WRS[*] boundary zones.	Churchill Lake, Wapesi Lake and Abamasagi batholiths; Wenasaga Lake, Sharp Lake and Root Bay (Lake St. Joseph) stocks.
Massive or foliated and/or lineated; locally cataclastic, green black; metamorphic textures dominant, medium and coarse grained; colour index 60 to 95.	Unknown	Cu-Ni-Co-Pd-Pt in gabbro-pyroxenite-peridotite inclusions distributed in Werner Lake fault zone.	Broadcast Lake and Conifer stocks; tectonic inclusions in peraluminous granites of eastern Lac Seul area.

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[*] *Abbreviations: TTG—trondhjemite–tonalite–granodiorite; ERS—English River Subprovince; WRS—Winnipeg River Subprovince; USP—Uchi Subprovince.*

and exhibit a considerably greater array and wider compositional span of internal intrusive phases than the uniform type (up to 10), as documented by Breaks and Bond (in prep.). These masses are broadly conformable with structural trends within the English River Subprovince and in the vicinity of the English River Subprovince–Winnipeg River Subprovince boundary and no deflection of the host rock planar fabric is discernible. Typically, these intrusive complexes are significantly larger than the uniform plutons. They can occur isolated within the English River Subprovince, as exemplified by the Bluffy Lake batholith (706 km^2), or along the English River Subprovince–Winnipeg River Subprovince boundary zone, as with the 2070 km^2 Chamberlain Narrows batholith (*see* Figure 7.2).

Both types of the tonalite-trondhjemite-granodiorite contain charnockitic rocks characterized by a brown to green-brown "greasy" lustre. These rocks contain orthopyroxene and are restricted to the granulite-grade zones delineated by Breaks et al. (1978). Examples (*see* Figures 7.2 and 7.3) include the Scrag Lake batholith, Gone Lake, Campfire Lake and Whitewater Lake stocks and numerous dikes. Orthopyroxene, commonly accompanied by lesser diopside, is normally disseminated throughout the rock. More conspicuous is the development of coarser-grained orthopyroxene in small patches of leucosome within an orthopyroxene tonalite host.

MODAL VARIATION

Modal compositions of the tonalite-trondhjemite-granodiorite suite are given in Figures 7.6b and 7.7b. Compositionally uniform plutons (e.g., Campfire Lake, McKenzie Bay, Goose Islands (Lac Seul) stocks; *see* Figure 7.2) exhibit limited variation in the QPK plot (*see* Figure 7.7b) and are mainly composed of rock varying from trondhjemite to tonalite in composition. The heterolithic, complex tonalite-trondhjemite-granodiorite intrusions (e.g., Chamberlain Narrows batholith) have not been studied in great petrographic detail, but the similarity of these rocks to the Sen Bay Stock on Lac Seul, for which petrographic data are available, makes it a likely representative of modal variation in this group. The Chamberlain Narrows batholith may be consanguineous with the nearby Sen Bay Stock (3008 Ma)

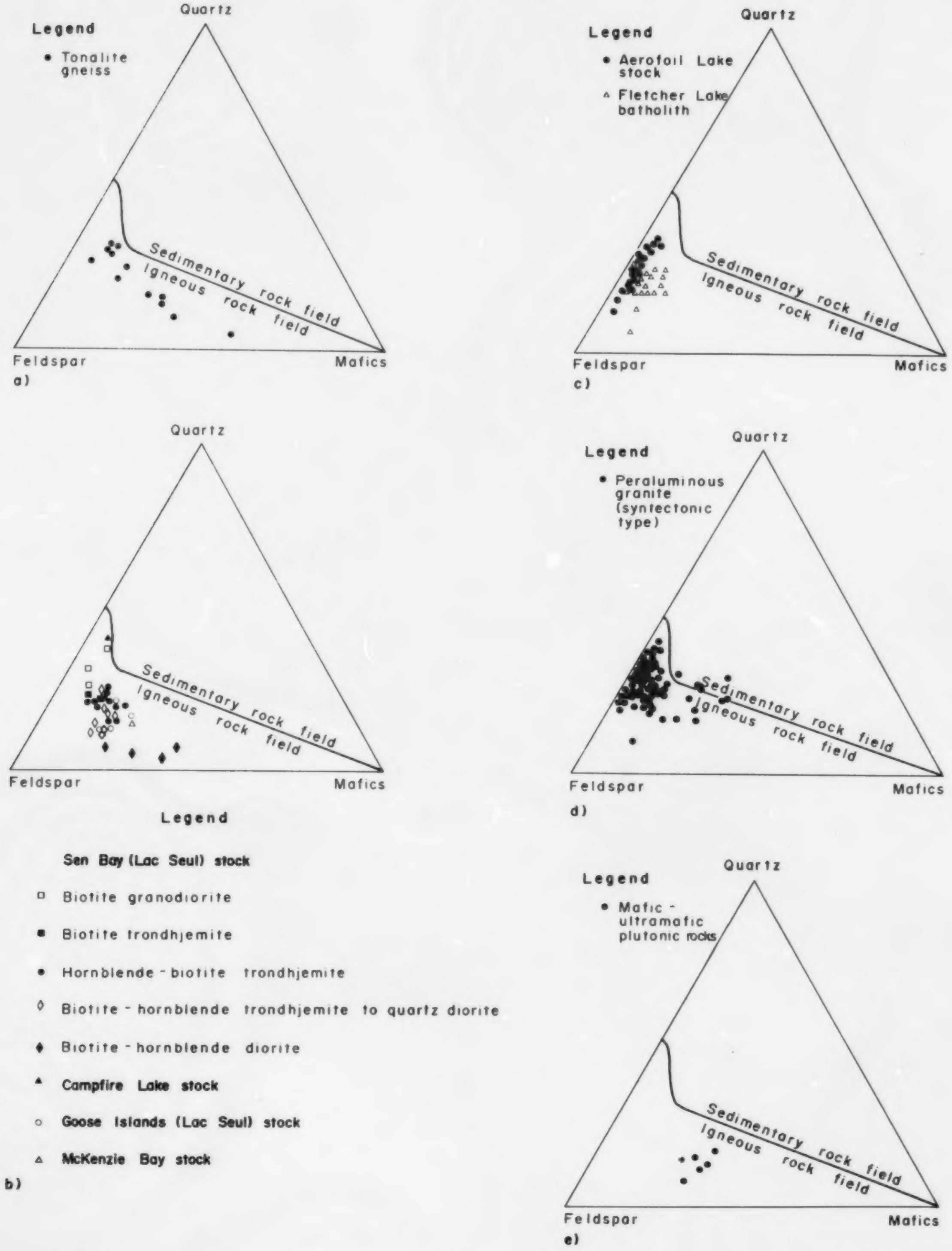


Figure 7.6. Modal variation of quartz, total feldspar and mafic minerals for plutonic suites of the English River Subprovince: **a)** gneissic tonalite suite; **b)** tonalite-trondhjemite-granodiorite suite; **c)** massive granodiorite to granite suite; **d)** peraluminous granite suite (syntectonic type); and **e)** mafic-ultramafic plutonic suite. Sedimentary and igneous rock compositional fields *modified from* van de Kamp and Beakhouse (1979).

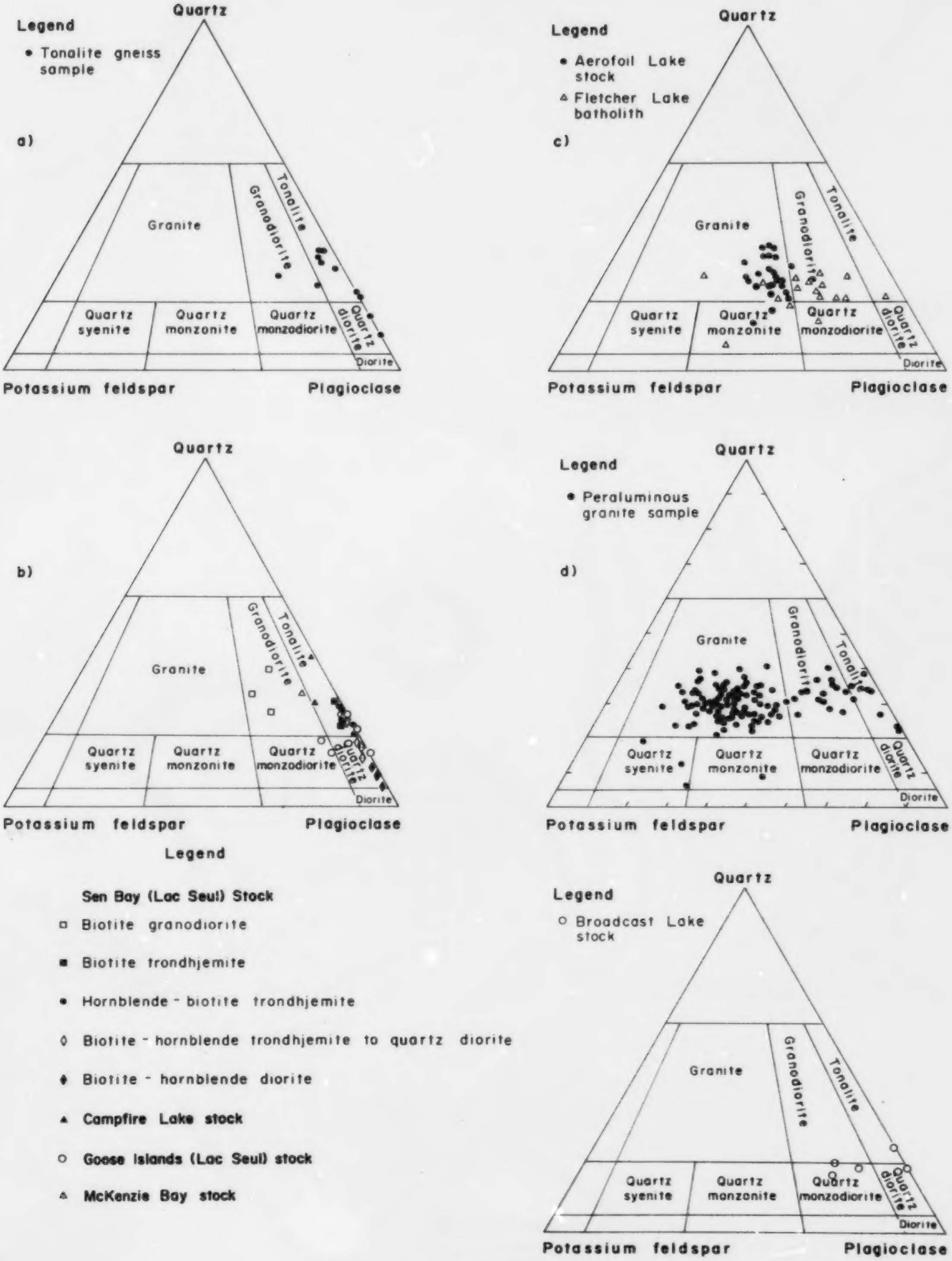


Figure 7.7. Modal variation of quartz, plagioclase and potassium feldspar for plutonic suites of the English River Subprovince: **a)** gneissic tonalite suite; **b)** tonalite-trondhjemite-granodiorite suite; **c)** massive granodiorite to granite suite (syntectonic type); **d)** peraluminous granite suite; and **e)** mafic-ultramafic plutonic suite.

based on a similarity of U-Pb zircon ages (Krogh et al. 1976; Corfu, unpublished data), and similarities in petrography and chemistry (Breaks and Bond, in prep.). Modal mineral data (Breaks and Bond, in prep.) for internal units from the Sen Bay Stock exhibit a continuum marked by a decrease in the colour index coupled with a decrease in the relative age in the generalized series: diorite-quartz, diorite-tonalite-trondhjemite-granodiorite (*see* Figure 7.6b). The QPK plot (*see* Figure 7.7b) reveals a variation marked by a wide range in plagioclase:quartz ratios; whereas potassium feldspar exhibits little variation except in the granodiorite plots which are separate from the main compositional field. Accessory minerals in the tonalite-trondhjemite-granodiorite suite include titanite, apatite, zircon, allanite, epidote, chlorite, muscovite and magnetite.

TIMING OF TONALITE-TRONDHJEMITE-GRANODIORITE SUITE MAGMATISM

Geochronologic work on the tonalite-trondhjemite-granodiorite suite in the English River Subprovince is limited. Interpretation of the data of Krogh et al. (1976), Wooden and Goodwin (1980), Wooden (NASA, personal communication, 1981) and Corfu (Royal Ontario Museum, personal communication, 1989) suggests an early period of magmatism: at *circa* 3.0 Ga along the English River Subprovince–Winnipeg River Subprovince boundary zone of the eastern Lac Seul region; and a later phase of tonalite-trondhjemite-granodiorite intrusive rocks within the metasedimentary migmatite belt at 2.65 to 2.698 Ga.

An early period of magmatism is represented by the greater than 3.0 billion-year-old Chamberlain Narrows batholith; a linear, east- to northeast-striking tonalitic complex forms an integral part of the English River Subprovince–Winnipeg River Subprovince boundary for approximately 65 km (*see* Figure 7.2). In the boundary, numerous slivers of amphibolitic mafic metavolcanic rocks (less than 0.5 km wide) and metasedimentary migmatites are interleaved with tonalitic rocks. The largest of the intrabatholithic metasedimentary belts is the Bear Narrows metasedimentary migmatite belt (*see* Figure 7.2) which mostly lies isolated in the pluton, 7 km south of the presently defined English River Subprovince–Winnipeg River Subprovince boundary. This belt consists principally of metasedimentary rocks migmatized at 2692 and 2668 Ma (Corfu, unpublished data). The metasedimentary rocks may be much younger than the Chamberlain Narrows batholith (Corfu, unpublished data), therefore, interleaving of supracrustal and tonalitic rocks occurred sometime between the 2760 million-year-old maximum age for sedimentation in the English River Subprovince (Krogh et al. 1976) and 2692 Ma when granulite-grade metamorphism of these metasedimentary rocks occurred (Corfu, unpublished data). This phase of English River Subprovince–Winnipeg River Subprovince boundary development occurred prior to the granulite-grade metamorphism, as is evident from the superimposition of the eastern Lac Seul granulite zone upon both subprovinces (*see* Figure 7.3).

A second period of tonalite-trondhjemite-granodiorite magmatism, approximately 300 my later than the *circa* 3 billion-year-old emplacement age of the Chamberlain Narrows batholith, forms an integral part of the tectono-magmatic evolution of the metasedimentary migmatite mass which dominates the English River Subprovince. This period is represented by both tonalite-trondhjemite-granodiorite suite types, and includes the Bluffy Lake batholith and Whitewater Lake (*see* Figure 7.2) and Adamhay Lake stocks (*see* Figure 7.9). This magmatic activity was coeval with the 2698 million-year-old syntectonic period of massive granodiorite to granite suite plutonism. Dated magmatic events younger than the second stage of tonalite-trondhjemite-granodiorite magmatism include *circa* 2668 million-year-old peraluminous S-type granite dikes and late dikes of the massive granodiorite to granite suite, mainly emplaced during the period 2560±40 to 2660±20 Ma (Krogh et al. 1976).

Massive Granodiorite to Granite Suite

This suite, forming only 6% of the English River Subprovince, was emplaced between 2698 and 2560±40 Ma, and thus overlaps and outlasts the peraluminous granite and tonalite-trondhjemite-granodiorite suite magmatism (*see* Table 7.2). Even though most of the rocks in this suite were generated during the syntectonic period at about 2698 Ma, 3 episodes of emplacement during the *circa* 2.7 billion-year-old Kenoran orogenic event have been recognized (Breaks and Bond, in prep.): syntectonic stocks and batholiths, late tectonic dikes and small masses, and posttectonic pegmatitic granite dikes which may locally be internally layered with aplite.

SYNTECTONIC GRANITOIDS

The sole large-scale representative of the syntectonic magmatic event mentioned previously is the 380 km^2 Fletcher Lake batholith (*see* Figure 7.2). Equivalent intrusions in the Winnipeg River Subprovince include the Lount Lake batholith (Breaks et al. 1978; Beakhouse 1983, this volume), dated at 2701.9^{+3}_{-9} Ma (Beakhouse and McNutt 1991), and the Cliff Lake granodiorite (Westerman 1977; Beakhouse 1983, this volume), dated at 2756 Ma (Corfu 1988). The massive to weakly foliated Fletcher Lake batholith dominantly consists of medium- to coarse-grained, potassium feldspar megacrystic granite and granodiorite with a local, weak rapakivi texture (*see* Figure 7.6c). A significant proportion of the batholith is charnockitic in composition (Breaks 1988). Small oval plutons of equigranular biotite-granite to quartz monzonite are further examples of the syntectonic massive granodiorite to granite suite (*see* Figures 7.6c and 7.7c). These rocks are far less abundant than the Fletcher Lake type, they exhibit a more regional distribution, such as the 1.2 by 3.5 km Aerofoil Lake stock (Breaks et al. 1976; Makepeace 1976). Emplacement of the homogeneous biotite granite has produced a concordant deflection of the banding in the host metasedimentary migmatite.

LATE TECTONIC GRANITOIDS

Examples of the late tectonic episode of massive granodiorite to granite suite magmatism in the English River Subprovince have been recognized in only a few localities. Mildly deformed, coarse-grained, magnetite-biotite granite was emplaced concordant to banding of metatexite in the Bear Narrows metasedimentary migmatite belt. This granitoid rock was emplaced at 2660±20 Ma (Krogh et al. 1976). Significantly, petrographically similar rocks in the Chamberlain Narrows (Lac Seul) area are the youngest that contain orthopyroxene. The massive granodiorite to granite suite rocks of the late tectonic period therefore represent a minimum age for static granulite-grade metamorphism in the eastern Lac Seul area (Breaks 1988).

POSTTECTONIC GRANITOIDS

Undeformed, coarse-grained to pegmatitic granite dikes represent the final stage of igneous activity in the English River Subprovince. These ubiquitous rocks were emplaced between 2660±20 and 2560±40 Ma (Krogh et al. 1976) and crosscut metasedimentary migmatite, the peraluminous granite suite and the youngest members of the tonalite-trondhjemite-granodiorite plutonic suite.

Peraluminous Granite Suite

Originally mapped as homogeneous and inhomogeneous diatexites, these rocks compose about 10% of the English River Subprovince (*see* Table 7.2). They were derived by regional anatexis of clastic metasedimentary rocks under low pressure, high-grade metamorphic conditions, during the Kenoran orogenic event (Breaks et al. 1978; Breaks 1982) at about 2681±20 Ma (Krogh et al. 1976). Emplacement of the extensive S-type peraluminous granite masses, at *circa* 2668 Ma, postdated the main regional phase of granulite-grade metamorphism. This relationship was established in an inclusion-rich part of the Churchill Lake batholith (*see* Figure 7.2) where orthopyroxene-bearing mafic blocks exhibit a strong foliation which developed during the dynamic phase of granulite-grade metamorphism. Peripheral parts of the blocks are overprinted by narrow orthopyroxene-rich selvages, formed by metasomatic reaction with the orthopyroxene-biotite granite host (Figure 7.8a).

Peraluminous granitic rocks are widespread in the English River Subprovince, accounting for a cumulative area of approximately 5000 km^2 between the Ontario–Manitoba boundary and long. 86°W (*see* Figure 7.2). Most of the suite is contained within 3 large bodies: the Churchill Lake batholith (2200 km^2), Wapesi Lake batholith (635 km^2) and Abamasagi batholith (1550 km^2) (*see* Figure 7.2). Dikes, stocks and small batholiths of this suite, commonly rich in muscovite and black tourmaline and geochemically fractionated, are intrusive on a regional scale into low- and medium-grade supracrustal rocks along the subprovince boundary zones. Plutons of this suite also occur in greenstone belts of the Uchi Subprovince, bordering the English River Subprovince, including the Papaonga Lake stock (Breaks et al. 1976) and the Twiname stock (Kay and Stott 1985) (*see* Figure 7.2) and in adjacent granitoid zones of the Wabigoon Subprovince (Smoothrock Lake stock; Sage et al. 1974).

Peraluminous granites (*see* Figures 7.6d and 7.7d) exhibit a ubiquitous, white weathered surface and possess a distinctive array of accessory minerals. These include the common presence of 2 micas (biotite and muscovite), coupled with any, or all of, garnet, cordierite, sillimanite and tourmaline. Of rarer occurrence are beryl, topaz, dumortierite and andalusite. A further distinctive mineralogical attribute is the complete absence of hornblende. The combination of these minerals produces high values of the aluminum saturation index (ASI) of Zen (1986), relative to other granitic suites in the English River Subprovince. The ASI for the peraluminous granites ranges from 0.96 to 1.17 (Breaks and Bond, in prep.). The peraluminous granites are not universally two-mica rocks; most of those associated with the granulite-grade zones, (e.g., Churchill Lake batholith, *see* Figure 7.2), contain only biotite. Only those plutons occurring near subprovince boundary zones or those associated with northeast-striking shear systems (e.g., Wapesi Lake batholith, *see* Figure 7.2) contain muscovite as a widespread varietal mineral.

MAFIC PERALUMINOUS ROCKS

Peraluminous granitic rocks rich in mafic minerals compose a small but regionally dispersed component of this suite. These granodiorites and tonalites dominantly occur within, and are proximal to the granulite-grade zones and are typically developed in units originally mapped as inhomogeneous diatexites (Breaks et al. 1978). The mafic peraluminous rocks have a higher colour index and less variation in accessory mineralogy than members of the tonalite-trondhjemite-granodiorite suite. This renders such rocks grossly similar to, and therefore difficult to distinguish from, leucocratic and mesocratic members of the tonalite-trondhjemite-granodiorite suite. A useful modal characteristic is the 15 to 30% quartz content, and the colour index typically exceeding 20. In contrast, an increasing colour index in members of the tonalite-trondhjemite-granodiorite suite correlates with a decreasing quartz content in this suite (*see* Figure 7.7d).

SYENITIC PERALUMINOUS ROCKS

Peraluminous quartz syenite and quartz monzonite are rare variants of the peraluminous granite suite (*see* Figure 7.6d), and are mainly restricted to the eastern Lac Seul area. Similar rocks may be present on Ogoki and Kayedon lakes (*see* Figure 7.2), where Kindle (1932, p.79) reported syenite and hornblendite dikes. Most syenitic rocks occur as diffuse zones within peraluminous granite-granodiorite or as coarse-grained to pegmatitic, orthopyroxene-biotite and biotite-garnet granitoid dikes that crosscut metatexitic metasedimentary migmatite. The diffuse zones may represent feldspar-cumulate zones (e.g., Anenimus Bay, eastern Lac Seul, *see* Figure 7.2). The origin of syenitic dikes is equivocal; they mainly postdate the initial migmatization

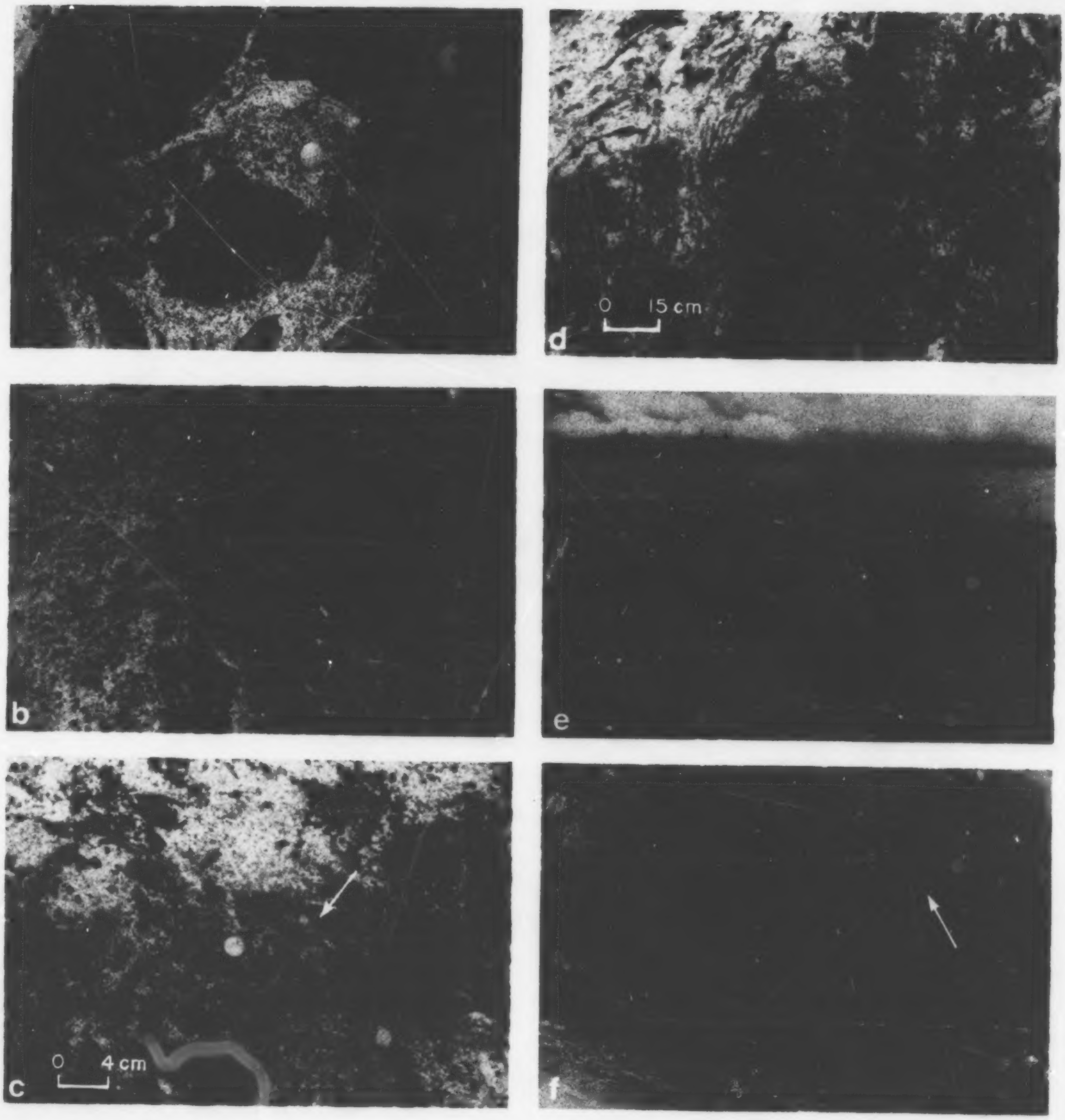


Figure 7.8. **a)** Heterolithic breccia-rich zone of orthopyroxene-biotite granite of the Churchill Lake batholith. The strongly foliated mafic fragments are subtly rimmed by a biotite-orthopyroxene selvage (arrow); Brisco Inlet, eastern Lac Seul. Coin in the centre of the photo is 2.5 cm in diameter; **b)** peraluminous granite with cordierite-quartz clots; Treelined Lake. Pencil is approximately 18 cm long; **c)** cordierite-quartz clot (arrow) in the process of separation from pelitic inclusion; Treelined Lake; **d)** inclusion of hornblende + diopside + orthopyroxene metaultramafic rock in inhomogeneous peraluminous granite; Treelined Lake; **e)** orthopyroxene-bearing amphibolite dike in foliated to gneissic tonalite of the Sen Bay Stock; southeast of Bear Narrows; and **f)** D_2 fold in metatexite which has refolded a D_1 isocline (arrow); near Conifer Lake. Coin in the centre of the photo is 2.5 cm in diameter.

event, but they could be the result of a second stage of anatexis imposed upon a quartz-depleted restite remaining after initial anatexis (cf., Clemens and Vielzeuf 1987).

PERALUMINOUS GRANITOID PEGMATITES

Chemically evolved derivatives of the peraluminous granite suite, termed "fertile granites" by Cerny and Meintzer (1988), occur as small granitoid plutons and pegmatites in lower grade greenstone belts of the Wabigoon Subprovince and Uchi Subprovince (*see* Figure 7.2). They are located immediately adjacent to the metasedimentary migmatites of the English River Subprovince (e.g., Root Lake; Breaks and Bond, in prep.), Twiname Bay (Kay and Stott 1985), and Separation Lake (Breaks and Bond, in prep.). These peraluminous granites are most likely to be the parent of the rare-element pegmatites enriched in beryllium, lithium, cesium, tantalum and rubidium which are characteristic of these boundary zones (Table 7.3). Pegmatite dikes, containing beryl, topaz, cassiterite, muscovite, tourmaline and garnet in the Linklater Lake area, are possibly apophyses of the Abamasagi batholith (*see* Figure 7.2).

CLOTTY PERALUMINOUS GRANITES

Peraluminous granites with a characteristic clotty texture are a petrographically interesting textural variant (*see* Figure 7.8b) and bears on the petrogenesis of the peraluminous granite suite. These rocks are regionally distributed in the English River Subprovince over a strike length of at least 240 km (e.g., Souter Lake in the east to Treelined Lake in the west). The best known and most accessible locality, at Perrault Lake, has been described by Morin (1970) and Morin and Turnock (1975).

Two petrographic types have been distinguished: simple clots, commonly composed of graphic intergrowths of cordierite and quartz; and complex clots composed of a more mafic assemblage of sillimanite + cordierite + biotite ± garnet ± plagioclase.

A restite origin for the biotite-rich clots may have resulted from partial melting of a pelite protolith (Morin and Turnock 1975; Beakhouse 1983; Breaks and Bond, in prep.). This mechanism of formation would certainly explain the restriction of the clots to a specific granite suite and the strong spatial link to migmatized high-grade metasedimentary rocks in the English River Subprovince. Supporting evidence for such an origin occurs in migmatite zones which reveal a transition for metatexite to inhomogeneous peraluminous granite (e.g., Treelined Lake, *see* Figure 7.8c).

A metasomatic origin for the clotty granites was first proposed by Cerny et al. (1967) to account for cordierite-quartz intergrowths in granitic pegmatites. Field evidence locally supports a metasomatic origin for cordierite-quartz clots in the English River Subprovince in the upper Root River area (*see* Figure 7.2), where linear segregations of cordierite-quartz are elongated parallel to the D_2 minor fold axes, and have overprinted the leucosome-mesosome interface in metatexite metasedimentary migmatite.

Gabbro, Diorite and Ultramafic Plutonic Suite

Examples of this plutonic suite compose less than 1% of the English River Subprovince (*see* Table 7.2 and Figure 7.2). Considerable petrographic diversity is displayed in the form of 3 compositional spectra (Breaks and Bond, in prep.); these are listed in order of decreasing abundance: diorite-quartz diorite- (minor gabbro) plutons; ultramafic-mafic intrusive complexes; dikes and derived tectonic inclusions (gabbro-pyroxenite-peridotite); and amphibolite dikes (*see* Figures 7.6e and 7.7e). These rocks range in area from stocks such as the Broadcast Lake stock, (*see* Figure 7.2) to inclusions of less than 1 m^2.

DIORITE-GABBRO PLUTONS

Two modes of occurrence have been documented (Breaks and Bond, in prep.): as isolated intrusive bodies and derivative inclusions; and bodies spatially and probably genetically associated with the tonalite-trondhjemite-granodiorite suite.

Isolated intrusions are the predominant members of the diorite-gabbro suite. For example, the Broadcast Lake stock, (*see* Figure 7.2; Breaks and Bond 1976; Desnoyers 1977), is a 4 by 32 km inequant mass of metamorphosed, massive to strongly foliated quartz diorite situated along the English River Subprovince–Winnipeg River Subprovince boundary. It intrudes Winnipeg River Subprovince tonalitic gneisses which are probably equivalent to the 3170^{+20}_{-5} million-year-old Cedar Lake gneisses (Corfu 1988), and post-dates anatexis that occurred between 2692 and 2668 Ma (Corfu, unpublished data) in English River Subprovince metasedimentary migmatites.

Rocks having a composition of gabbro, diorite and quartz diorite form an early minor component of tonalite-trondhjemite-granodiorite suite plutons. Examples of such rocks include, hornblende-biotite diorite in the Celt Lake area of the Bluffy Lake batholith, and diorite units in the 2697 million-year-old Adamhay Lake stock and the greater than 3.0 billion-year-old Chamberlain Narrows batholith (*see* Figure 7.2).

GABBRO-PYROXENITE-PERIDOTITE PLUTONS

Plutonic rocks forming gabbro-pyroxenite-peridotite intrusive complexes are rare in the English River Subprovince and are confined to 3 areas. These areas generally coincide with zones of granulite-grade metamorphism: areas adjacent to Conifer–Sumach lakes; areas adjacent to Reynar, Rex, Treelined and Gone lakes; and the eastern Lac Seul region (*see* Figure 7.2). Ultramafic inclusions, a metre or less in diameter, are widespread in the English River Subprovince.

Ultramafic to gabbroic rocks consist of clinopyroxene-, orthopyroxene- and olivine-bearing pyroxenite and

Table 7.3. Examples of mineralization types in the English River Subprovince; typical concentrations, mineral and elemental associations and source of reference are listed (all locations are shown on Figure .2).

Deposit Type	Examples	Grade	Mineralogical and Geochemical Association	References
I Rare-Element Pegmatites Beryl type	Sandy Creek	139 ppm Be	Beryl, tourmaline, apatite, purpurite ([Mn^{+3}, Fe^{+3}]PO_4) (Be-B-P-Nb>Ta)	Chisholm (1948)
Linklater Lake	~1% Sn		Beryl, cassiterite, tourmaline, microcline, topaz (Be-Sn-B)	Mulligan (1960)
Albite-spodumene type	Root Lake	1.3% Li_2O	Spodumene, beryl, tourmaline, microcline, lepidolite, columbite–tantalite (Li-Rb-Be-Ta≥Nb)	Mulligan (1960) Breaks and Bond (in prep.)
Complex spodumene type	Lilypad Lake	1.07% Li_2O 2.54% Cs_2O	Petalite, spodumene, rubellite, lepidolite, fluorite, etc. (Li-Cs-Rb-Be-F-Ta>Nb)	Prest (1944) Mulligan (1965)
II Radioactive Element-Enriched Pegmatites	Sydney Lake	0.1% U_3O_8	β-uranophane, garnet, biotite (U>Th)	Breaks and Bond (in prep.)
	Umfreville Lake	52 ppm U, 108 ppm Th	Uraninite, β-uranophane, monazite, molybdenite	Breaks (1982) Breaks and Bond (in prep.)
III Banded Iron Formation	Griffith Mine Skibi Lake	26% Fe 26% Fe	Magnetite, quartz, minor sulphide Magnetite, quartz, minor sulphide	Shklanka (1968) Shklanka (1968)
IV Sulphide Minerals (in ultramafic-mafic association)	Werner Lake	2% Co	Cobaltite, chalcopyrite, etc.	Carlson (1957) Lawson and Zuberec (1987)
	Gordon Lake Mine	~1 ppm Pd	Chalcopyrite, pyrrhotite, pyrite	Carlson (1957) Hailstone and Blackburn (1987) Lawson and Zuberec (1987)
V Volcanogenic Massive Sulphide	Kapikotongwa Lake	Zn <4.70% Cu <1.81% Pb <2.86%	Sphalerite, galena, pyrite, chalcopyrite (Zn-Pb-Cu-Cr-Ag-Au)	Bond and Foster (1981a, 1981b)

peridotite, and associated gabbroic rocks (Carlson 1957; Panagapko 1976; Breaks and Bond, in prep.).

The largest single mass of ultramafic to gabbroic rocks is the Conifer stock (*see* Figure 7.2), a circular complex 1200 m across, (Panagapko 1976). This complex lies in the centre of the northeast arm of a large peraluminous mass composed of biotite + garnet ± sillimanite ± cordierite-bearing granite delineated by Breaks, Bond, McWilliams et al. (1975). This granite intrudes the core of the ultramafic-mafic pluton (Panagapko 1976).

The Conifer stock consists of metamorphosed, massive, medium- to coarse-grained ultramafic rocks which are chemically similar to pyroxenite, and to a lesser extent to subordinate melagabbro and meladiorite. The ultramafic rocks form a 280 m wide outer ring and are composed of hypersthene + biotite + hornblende ± plagioclase ± opaque iron-oxide. The associated meladiorite, composed of hornblende + augite + andesine + biotite ± olivine ± hypersthene, has a lower colour index (75 to 85) than pyroxenite. Foliations developed along the inner and outer granite contacts have a cataclastic appearance (Panagapko 1976), with foliation consistently dipping inwards between 30 and 55, which suggests a funnel-like morphology for this body.

Ultramafic inclusions are principally found in the peraluminous granite suite (*see* Figure 7.8d) and more rarely are found in tonalite-trondhjemite-granodiorite suite rocks of charnockitic affinity (e.g., Gone Lake stock, *see* Figure 7.2). The inclusions are composed entirely of a metamorphic biotite + plagioclase + diopside + orthopyroxene assemblage. On a regional basis, inclusions distributed in peraluminous granite characteristically contain a biotite-rich reaction rim which developed via metasomatic interchange with former fluids of its potassium rich granite host. Similar selvages of biotite-hornblende schist have been documented in the Rex–Werner lakes area around ultramafic inclusions measuring 20 by 65 m (*see* Figure 7.2; Carlson 1957; Lawson and Zuberec 1987). Tectonized remnants of ultramafic rock, preserving rare D_1 fold hinges (e.g., Chamberlain Narrows, Lac Seul) are found in a metatexite metasedimentary migmatite. Perhaps some of these inclusions in the S-type granites may originally have been emplaced as dikes or tectonic slivers prior to the anatectic and granulite-grade metamorphic events.

AMPHIBOLITE DIKES

Relicts of a north-striking, mafic dike swarm, emplaced before granulite-grade metamorphism at 2692 Ma, are represented by rocks of plagioclase + hornblende ± quartz ± biotite ± orthopyroxene ± diopside composition. These relicts occur principally in the eastern Lac Seul region as 0.3 to 3 m wide dikes that intrude most major rock units (*see* Figure 7.8e). In places, the amphibolites form linear "trains" of deformed enclaves.

MINERALIZATION

Introduction

The English River Subprovince has a low economic mineral potential relative to that of the greenstone-dominated neighbours. Mineralization types include (*see* Table 7.3): rare element and radioactive element-enriched pegmatites; iron formation deposits; cobalt-copper-nickel-platinum group metals; volcanogenic massive sulphide deposits; and rare gold. Early interest in deposits centred on cobalt and iron, and culminated in the opening of the Griffith Mine in 1968 (Shklanka 1968, 1970). Cobalt was discovered at Werner Lake in 1920 and mined there from 1920 to 1944 (Carlson 1957). Copper, nickel and platinum group elements concentrations were found in that area also and led to production at the Gordon Lake Mine between 1963 and 1969 (Hailstone and Blackburn 1987).

Pegmatite-Hosted Deposits

A significant feature of the English River Subprovince is the linear, regional distribution of rare-element pegmatites within the low- to medium-grade metamorphic rocks along the northern and southern boundary zones (Breaks et al. 1978; Cerny and Trueman 1978; Breaks et al. 1985).

The rare-element pegmatites have a diverse mineralogy and chemistry (*see* Table 7.3) and correspond to beryl, albite spodumene, and complex spodumene types (Beus 1960; Rudenko et al. 1975; Cerny 1988). The English River Subprovince pegmatites are genetically related to small masses of less than 10 km² area, of chemically fractionated, peraluminous, S-type granites (Breaks 1989, 1990).

Radioactive element-enriched pegmatites are mainly situated west of long. 93°W (*see* Figure 7.2). Concentrations of uranium and thorium are found in 2 distinct granite types (Breaks 1982): peraluminous S-type pegmatitic granites, and pegmatites of the massive granodiorite to granite suite. Levels of uranium oxide are presently subeconomic at 0.07%.

Iron Formation Deposits

Banded iron formation of turbiditic association is the most extensive deposit type in the English River Subprovince (*see* Table 7.3). Numerous prospects have been evaluated, these vary in size and grade from 21 to 335 Mt and 21.6 to 31.2% Fe (Shklanka 1968, 1970).

Sulphide and Gold Deposits

Examples of copper, nickel, cobalt, palladium and platinum mineralization (*see* Table 7.3) are confined to blocks of mafic to ultramafic plutonic rocks within the Werner Lake fault zone (Carlson 1957; Hailstone and Blackburn 1987; Lawson and Zuberec 1987).

Volcanogenic massive sulphide deposits are scarce in the English River Subprovince (*see* Table 7.3), known only within the Melchett Lake metavolcanic-metasedimentary and Separation Lake greenstone belts. Local fracture fillings and disseminations of sphalerite and pyrite, are accompanied by galena and chalcopyrite and rare amounts of native gold in a 700 m thick felsic pyroclastic host rock (Bond 1979; Bond and Foster 1981a, 1981b). High- grade metamorphosed massive sulphide mineralization occurs in the Moose Lake–Bury Lake area (Hudec 1965). Gold is also found within quartz veins near Saga Lake within the Melchett Lake metavolcanic-metasedimentary belt (Thurston and Carter 1970; Riley et al. 1971).

STRUCTURAL GEOLOGY

Introduction

The structural analysis of the English River Subprovince presented in this section is derived largely from regional mapping in Ontario undertaken by Breaks et al. (1978), Breaks and Bond (in prep.), as well as localized studies (Westerman 1977; Stone 1981; Sanborn-Barrie 1988; Williams 1988). The structural database is such that it is not clear to what extent all phases of deformation are present in any one area. Within this large area of variably metamorphosed rocks which have contrasting mechanical properties, the distinction and relative timing of deformation phases using structural style, orientation and overprinting relationships is problematic. This section does not assign every deformation structure to a deformation phase, and recognizes that there are regional problems of fold phase correlation and kinematic significance.

On a regional scale, the English River Subprovince is dominated by an east-striking mineral foliation that dips south to southwestward at angles of 15° to 90°, but is generally subvertical. Typically, mineral foliation and compositional banding are coplanar in the predominant metasedimentary migmatite. Around some granitic plutons, there is a localized concentric deflection of the foliation (e.g., Campfire Lake stock, *see* Figure 7.2). There are also local curvilinear expressions of macroscopic folds and domal structures (e.g., Churchill Lake area, *see* Figure 7.2), or, within 1 to 2 km of obliquely striking faults, the planar fabric has been rotated into concordance with such zones (e.g., Miniss River Fault, Breaks et al. 1979a).

Mineral lineations associated with the regional mineral foliation are variably developed, and are most intense within the subprovince boundary zones. Over most of the English River Subprovince, mineral lineations trend eastward at shallow to moderate plunges of 20° to 70°. In contrast, Sanborn-Barrie (1988, p.104) documented a 5 to 12 km wide domain along the English River Subprovince–Winnipeg River Subprovince boundary, containing lineations which trend in a westerly direction at 55° to 70°.

Structural Development

Rocks within the metasedimentary migmatite-dominated English River Subprovince were subjected to 3 phases of folding, and a protracted phase of kink-style folding, shearing, fracturing and faulting (McRitchie and Weber 1971; Breaks et al. 1978; Breaks and Bond, in prep; Westerman 1977; Stone 1981; Sanborn-Barrie 1988; Williams 1988). These phases will be described herein in the order of their development. Whereas, McRitchie and Weber (1971) chose to assign a separate phase of deformation to their 3 final groups of faults and shear zones, such a distinction cannot be verified by the scant data recorded in Ontario. A distinction between the planar tectonic elements S_1 and S_2 cannot be made with confidence on a regional scale. The dominant, regional, foliation is best regarded as a composite of bedding (S_0), S_1 and S_2. However, in the high-grade metasedimentary migmatite terrane, rare exposures such as those at Aerofoil Lake (*see* Figure 7.2), display bedding lamination that is coplanar with S_1. Both of these have been discordantly overprinted by an S_2 foliation axial planar to close D_2 folds. In the better preserved low- to medium-grade metamorphic terrane, S_0 is generally parallel to a later foliation S_1 and/or S_2.

D_1 DEFORMATION

The earliest deformation is recognized by rare mesoscopic, shallow-plunging, tight to isoclinal folds of bedding in rocks which have a low metamorphic grade in the northern part of the English River Subprovince. Fold axes have a variable trend, but are dominantly parallel to the boundaries of the English River Subprovince. Macroscopic D_1 folds have not been recognized, but were probably responsible for the juxtaposition of *circa* 3 billion-year-old tonalite against metasedimentary rocks in the southern part of the subprovince. Large-scale, early, recumbent folds were inferred to have developed in the northern boundary zone by Thurston and Breaks (1978).

Rootless intrafolial folds which were enveloped by the regional D_2 planar fabric could represent relicts of D_1 deformation. The relative chronology of D_1 and D_2 folds has been established by interference patterns in a few scattered localities, (e.g., Wabaskang Lake, Westerman 1977; Churchill Lake and near Conifer Lake, *see* Figure 7.8f). At the Churchill Lake locality, wacke exhibits D_1 isoclinal folding of a thin arkose bed; the shallow southward dip of the D_1 axial surface is folded by upright, more open, D_2 folds.

The development of an S_1 mineral foliation parallel to bedding in low-grade metamorphic rocks, and the enhancement of compositional banding in the migmatitic metasedimentary rocks, has been inferred by Westerman (1977). The S_1 mineral foliation, formed during D_1 deformation, has only been locally established because further strain and metamorphic coarsening of the matrix accompanying the D_2 event has obliterated much of the D_1 structure. The S_1 fabric is locally preserved in 3 ways: in D_1 fold hinges; as inclusion trains within M_1 porphyroblasts of staurolite, biotite and andalusite (these minerals are posttectonic with respect to D_1); and as a bedding-parallel foliation preserved in metasedimentary inclusions within peraluminous granites having a D_2 foliation. Sanborn-Barrie (1988) recognized an S_1 gneissosity oriented oblique to the English River Subprovince–Winnipeg River Subprovince boundary over a 260 km strike length between Umfreville Lake and eastern Lac Seul (*see* Figure 7.2). Contrasting attitudes of S_1 in the English

River Subprovince and in the adjacent Winnipeg River Subprovince (Beakhouse, this volume) indicate that D_1 may have developed when these 2 subprovinces were physically separate (Sanborn-Barrie 1988).

D_2 DEFORMATION

This regional deformation event is characterized by commonly disharmonic, tight to close, symmetric and asymmetric folds with Z-styles dominant over S-styles, and moderate east- to southeast-plunges. The folds have an associated strongly developed axial planar fabric (S_2) that parallels the regionally developed east-striking, steeply dipping foliation that is concordant to the subprovince boundaries and is generally considered to be coplanar with S_1. Axial planar foliations of D_1 and D_2 folds are usually indistinguishable except where the foliations are discordant at angles of greater than 10. Macroscopic Z-style asymmetric D_2 folding of the Separation Lake greenstone belt (*see* Figure 7.2) and an associated regionally developed axial planar S_2 schistosity (Sanborn-Barrie 1988, p.102) at the English River Subprovince–Winnipeg River Subprovince boundary is one of the few examples of large-scale D_2 folding. Macroscopic D_2 antiforms and synforms also occur between Churchill and Miniss lakes (*see* Figure 7.2; Breaks et al. 1976; Williams 1977) and in the Burntrock Lake area (Sage et al. 1974).

S_2 fabric apparently controlled emplacement, orientation and timing of mobile peraluminous granite leucosome in many parts of the English River Subprovince.

D_2 deformation may have developed as a response to compression across the subprovince (Westerman 1977). Sanborn-Barrie (1988) documented a predominance of S-style asymmetric folds and west-plunging lineations at the English River Subprovince–Winnipeg River Subprovince boundary and interpreted them to be supportive of oblique sinistral overthrust shearing of the Winnipeg River Subprovince tonalitic rocks over English River Subprovince metasedimentary migmatites. D_2 folds have also effected local sinistral rotation of the M_1 porphyroblasts.

D_3 DEFORMATION

Sporadic kink-like folds with steeply plunging axes and both Z- and S-asymmetry occur in metasedimentary migmatite within the English River Subprovince and are assigned to D_3 (Westerman 1977). D_3 folds exhibit a variable style, producing asymmetric, close to open, concentric S-folds which plunge steeply northwest with northwest-striking axial planes and nonpenetrative planar fabric (Westerman 1977, p.205, *nb*, Figure 4–10 is of a Z-style kink fold at Wabaskang Lake). The D_3 deformation event was attributed by McRitchie and Weber (1971) to the waning stages of the D_2 event. Leucosome development is locally associated with these folds; however, the amount of leucosomes is insignificant relative to D_2 leucosomes.

In contrast, Sanborn-Barrie (1988) assigned narrow dextral sense, strike-slip shear zones and major dextral sense shearing along or parallel to the English River Subprovince–Winnipeg River Subprovince boundary to "Sanborn-Barrie's" D_3 event. It seems likely that localized and contrasting styles of deformation were able to take place within the English River Subprovince with the possibility that the D_3 of Sanborn-Barrie (1988) may be equivalent to one of the later, fault-related folding deformations along the northern edge of the English River Subprovince (e.g., Stone 1981).

BRITTLE AND SEMIBRITTLE DEFORMATION

There is scant information recorded regarding the late stages of inhomogeneously distributed structures and nonpenetratively developed deformation. However, deformation apparently became increasingly brittle in character and affected progressively smaller volumes of rock.

Kink Folding and Warping

In Manitoba, McRitchie and Weber (1971) attributed broad, highly schistose, curvilinear shear zones up to 1 km wide, to examples of a D_4 deformation. The generation of major shear zones, such as the Moose River Fault in Manitoba that becomes the Sydney Lake–Lake St. Joseph Fault in Ontario (Stone 1981; Williams 1988) is the first stage in the development of the English River Subprovince–Uchi Subprovince boundary.

Regional Z-warping, kinks with Z- and S-asymmetry, and fractures locally deflect mylonite zones and are spatially associated with them. In Manitoba, these warps and kinks have steeply dipping axial surfaces which strike northwards to northeastwards (McRitchie and Weber 1971).

Faulting

Three regional-scale faults of differing attitudes transect the English River Subprovince (*see* Figure 7.2): the east-striking Sydney Lake–Lake St. Joseph Fault (Breaks, Bond, McWilliams et al. 1975; Stone 1981; Williams 1988), the northeast-striking Miniss River Fault (Hudec 1965; Breaks and Bond 1977; Breaks et al. 1981) and the northwest- to west-striking Pashkokogan Fault Zone (Sage et al. 1974, Breaks et al. 1978). The time interval involved for the fault development in the English River Subprovince is unknown, but may be considerable (Osmani, this volume). In the English River Subprovince, fault systems represent the only tectonic deformation that has been observed to overprint the youngest members of the massive granodiorite to granite suite, implying a maximum age of 2701±1 Ma for the Sydney Lake–Lake St. Joseph Fault based upon the overprinting of this fault upon the Bamaji–Blackstone granite (G.M. Stott, Ontario Geological Survey and F. Corfu, Royal Ontario Museum, personal communications, 1989).

The Sydney Lake–Lake St. Joseph Fault is a 1 to 2 km wide zone of ductile deformation composed of protomylonite and mylonite developed from metatexite metasedimentary migmatites, peraluminous granites, metamorphosed felsic to intermediate intrusive rocks and metavolcanic units. Major lithologic units, such as the Pineneedle Lake pluton (*see* Figure 7.2) and Bee Lake belt (Stott and Corfu, this volume) can be traced into the fault zone, and the offset of these units indicates horizontal, dextral strike separation of at least 30 km (Stone 1981). The component of dip-slip associated with the latest period of transcurrent motion was calculated to be 2.2 to 2.6 km by Stone (1981), who examined the plunge variation of slickenside lineations and shear strain across the Sydney Lake–Lake St. Joseph Fault. A disparity between this estimate and a geobarometrically determined one of 6 km by Campion et al. (1986) may be accounted for by an earlier period of dip-slip motion. For most of its length, the Sydney Lake–Lake St. Joseph Fault marks an abrupt lithologic and metamorphic discontinuity between low- to medium-grade metavolcanic-metasedimentary assemblages of the Uchi Subprovince and high-grade metasedimentary migmatites and metaplutonic rocks of the English River Subprovince. The sharp lithologic change from felsic to intermediate granitic rocks in the Uchi Subprovince to metasedimentary migmatite of the English River Subprovince, is best exposed between Pakwash Lake (*see* Figure 7.2) and the eastern termination of the Bee Lake belt (Stott and Corfu, this volume). Brittle, dextral sense, strike-slip reactivation of earlier ductile motions on the Sydney Lake–Lake St. Joseph Fault locally produced small amounts of fault breccia and pseudotachylyte (Stone 1981; Williams 1988; Kamineni et al. 1990), which has been dated at 2183±74 Ma.

Sinistral faults, with northeast and northwest strikes, obliquely transect the English River Subprovince. One example is the northeast-trending Miniss River Fault that is traced for 135 km (Breaks and Bond 1977; Breaks et al. 1979a, 1979b, 1981). A possibly related fault system overprints the Wapesi Lake batholith (*see* Figure 7.2) with protomylonite and mylonite zones developed in the peraluminous granite. The fault system produces distinctive geophysical patterns for this part of the English River Subprovince which consists of a combined linear gravity and aeromagnetic low.

The east-southeast-striking Pashkokogan Fault Zone, like the Miniss River Fault, is in part oblique and, locally, parallel to the boundaries of the English River Subprovince. Along the English River Subprovince–Wabigoon Subprovince boundary, east-trending asymmetric Z-folds with shallow plunges and associated fish structures (Hanmer 1985) in highly strained metasedimentary metatexite are developed at least over a width of 3 km (Sutcliffe 1988) from which dextral, strike-slip motion can be inferred. Where the fault crosses tonalite-trondhjemite-granodiorite suite rocks, both ductile mylonites and fault breccias are developed.

STRUCTURAL DEVELOPMENT OF THE ENGLISH RIVER SUBPROVINCE–WINNIPEG RIVER SUBPROVINCE BOUNDARY

The interpreted tectonic interleaving of young migmatized metasedimentary rocks and significantly older tonalitic rocks documented for the eastern Lac Seul region, hinders precise definition of the English River Subprovince boundary with the Winnipeg River Subprovince. It appears that there is a continuum of rock types and structural style across the presently defined subprovince boundary in this region. The original boundary, probably tectonic, has been partly obliterated by younger granitoid intrusions, and has been folded about easterly trending axes. Similar relationships have been noted for the Wawa–Quetico subprovince boundary (*see* Williams, this volume).

Towards the boundary between Ontario and Manitoba, the English River Subprovince–Winnipeg River Subprovince boundary is marked by the Bird River–Separation Lake belt (*see* Beakhouse, this volume). The tectonic significance of these metavolcanic rocks is complicated by disagreement over whether they should be considered an integral part of one, or other, or neither of the 2 subprovinces. The 5 to 12 km wide supracrustal domain is characterized by lineations with a trend reverse to that normally seen (Sanborn-Barrie 1988). Structurally, therefore, there is justification for distinguishing the Bird River Subprovince from the adjacent subprovinces.

METAMORPHISM

This section reviews the nature of metamorphism in the English River Subprovince and its timing in relation to the deformation and migmatization events. An evaluation is made of the potential causes of this Kenoran (2.66 to 2.70 Ga) metamorphism which affected all rock types exclusive of Proterozoic diabase and the late orogenic to postorogenic members of the massive granodiorite to granite suite and of the peraluminous granite-granodiorite suite.

Patterns and Character of Regional Metamorphism

Prior to the regional mapping of Breaks et al. (1978), very little was known about metamorphic patterns in the English River Subprovince of Ontario. Transects revealed that the subprovince was generally composed of high-grade rocks, with local granulite-grade zones These zones decrease to low and medium grade along the English River Subprovince–Uchi Subprovince boundary and, more locally, decrease to low and medium grade along the Separation Lake greenstone belt at the English River Subprovince–Winnipeg River Subprovince boundary in western Ontario (Dwibedi 1968; Jones 1973; Trueman et al. 1975).

Thurston and Breaks (1978, p.58) originally proposed that the metamorphic pattern could be described in terms of a thermal anticline positioned over the combined English River Subprovince and Winnipeg River Subprovince. These authors also considered the apparent upward curvature of isograds in the sedimentary pile was due to tectonic thickening and subsequent erosion (Richardson 1970). The gross distribution (*see* Figure 7.3) of metamorphic zones, however, reveals an asymmetric pattern with respect to the English River Subprovince–Winnipeg River Subprovince boundary. There is deviation from this pattern where low- and medium-grade metavolcanic rocks of the Wabigoon Subprovince and Bird River Subprovince abut segments of the southern boundary with the English River Subprovince; there, the metamorphic pattern is nearly symmetric (Trueman et al. 1975; Thurston and Breaks 1978;

Breaks et al. 1978; Breaks 1988). A further deviation from the overall metamorphic asymmetry occurs in the medium-grade Melchett Lake metavolcanic-metasedimentary belt which lies completely isolated within high-grade metasedimentary migmatite and derived peraluminous granite (*see* Figure 7.3).

METAMORPHIC FACIES

The regional metamorphic facies series are commonly interrupted and telescoped by a component of vertical motion along the boundary fault zones. Abrupt changes in metamorphic grade occur across the Sydney Lake–Lake St. Joseph Fault in several localities. For example, in the Bee Lake–Rice Lake area, the fault juxtaposes medium-grade andalusite-bearing pelites and high-grade metasedimentary migmatite, a metamorphic discontinuity corresponding to an abrupt rise in temperature from 500° to 600°C and an increase in pressure from 0.4 to 0.6 GPa (Campion et al. 1986). Metamorphic discontinuities of similar magnitude can also be found in the Pashkokogan Lake and Pakwash Lake areas.

Relatively intact metamorphic facies series, containing all metamorphic zones recognized on a regional scale are preserved in only a few areas. For example, 6 metamorphic zones have been delineated between Lake St. Joseph and eastern Lac Seul (Breaks et al. 1978). These zones, as listed below, are mainly derived from the mineral assemblages developed in pelitic and wacke rocks. The mineral assemblage database is locally augmented by those developed in mafic rocks.

LOW GRADE	Chlorite-biotite zone
MEDIUM GRADE	Staurolite-chlorite-biotite zone Sillimanite-muscovite zone
HIGH GRADE	Sillimanite-potassium feldspar zone Cordierite-garnet-potassium feldspar zone Orthopyroxene zone

GRANULITE-GRADE METAMORPHISM

Low-pressure granulite-grade metamorphism, first described in the English River Subprovince by Breaks et al. (1978), occurs as oval nodes up to 110 km long by 25 km across, intermittently dispersed over a strike length of 500 km (*see* Figure 7.3). These zones represent apparent thermal culminations and tend to occur close to the English River Subprovince–Winnipeg River Subprovince boundary. Notably 3 such granulite-grade zones "overprint" this boundary. Workers such as Baumann et al. (1984) have characterized the granulite-grade metamorphism as being of medium-pressure type. However, the estimated pressure interval of Baumann et al. (1984) lies largely within the low-pressure granulite-grade field of Green and Ringwood (1967).

Recently, Breaks (1988) divided metamorphism in the eastern Lac Seul granulite zone into a dynamic (synkinematic) and a static (late kinematic to postkinematic) type, and documented their timing relative to structural and magmatic development. The dynamic type granulite-grade metamorphism is the most widespread (Figure 7.9) and is characterized by orthopyroxene developed in association with D_2 planar and linear fabrics, which locally envelop D_1 rootless intrafolial isoclines. Static granulite-grade metamorphism is less extensive and is characterized by a series of mineral assemblages containing orthopyroxene, biotite, plagioclase and iron-oxides. Field relationships between granulite-grade assemblages, magma intrusion and deformation indicate that 5 distinct metamorphic pulses occurred during the late kinematic to postkinematic history of the region; static granulite-grade metamorphism outlasted the emplacement of the tonalite-trondhjemite-granodiorite suite. The development of granulite-grade selvages has been documented around tonalitic, granitic, mafic and syenitic intrusive rocks in gash and boudin fillings (Breaks 1988). Infiltration of tonalitic and granite-granodiorite melts coexisting with a CO_2-rich fluid may have induced metasomatic fluid exchange, and consequent static granulite-grade metamorphism may have been induced by destabilization of hornblende and diopside and their replacement by orthopyroxene in the selvage zones.

CONDITIONS OF METAMORPHISM

Conditions of pressure and temperature during metamorphism have been estimated from comparison of English River Subprovince mineral assemblage data with experimentally determined conditions for relevant mineral reactions (Breaks et al. 1978; Thurston and Breaks 1978) and by thermobarometric determinations (Henke and Perkins 1983; Henke 1984; Chipera 1985; Percival et al. 1985; Campion et al. 1986; Chipera and Perkins 1988). These data reveal that the regional metamorphic character of the English River Subprovince favourably compares with the high-temperature, low-pressure intermediate series (Abukuma type) of Miyashiro (1961). The observed facies series (Figure 7.10) was generated by a metamorphic field (vertical) gradient between 35° and 50°C/km (Breaks et al. 1978).

Pressure conditions in the English River Subprovince correspond to Bathozone 2 of Carmichael (1978) and exhibit little regional variation (Henke 1984; Chipera 1985). Medium-pressure Barrovian metamorphic assemblages are virtually absent, which is also the case in the geologically similar Quetico Subprovince (Percival 1989; Tabor et al. 1989; Bauer and Tabor 1990; Williams, this volume). Kyanite-bearing rocks accompanied by sillimanite have only been recognized in altered felsic metavolcanic rocks of the Melchett Lake area (Bond and Foster 1981b). The local association of kyanite with cordierite and muscovite suggests load pressures exceeding 0.5 GPa at 550°C (*see* Figure 7.10).

Temperatures attained during granulite-grade metamorphism are estimated to have exceeded 770°C, with a load pressure range of 0.4 to 0.6 GPa (Breaks et al. 1978). Garnet-biotite thermometry and garnet-orthopyroxene-plagioclase-quartz barometry in the eastern Lac Seul granulite zone resulted in temperature estimates of 675° to 750°C and pressure estimates of 0.47±0.1 to 0.54±0.8 GPa (Baumann et al. 1984; Perkins and Chipera 1985;

Chipera and Perkins 1988). Similar values of 680° to 730°C and 0.4 to 0.5 GPa were obtained in the Whitewater–Mojikit lakes granulite zone by Digel (*in* Percival et al. 1985), using garnet + sillimanite + plagioclase + quartz assemblages in metasedimentary rocks and garnet + clinopyroxene + orthopyroxene + plagioclase + quartz in assemblages mafic rocks. The highest grade pelitic rocks encountered in the English River Subprovince contain hercynite enveloped by cordierite, suggesting temperatures of 750°C (Henke 1984). These data indicate that the presently exposed granulite-grade zones have been exhumed from at least a depth of 20 km .

In the lower grade marginal zones, amphibolite-facies metamorphic conditions fall into a range of 550° to 650°C and 0.3 to 0.45 GPa (Breaks et al. 1978, p.46) which overlap with the more precise thermobarometric determinations of 500° to 625°C and 0.4 GPa (Chipera 1985; Campion et al. 1986).

GEOCHRONOLOGY OF METAMORPHIC EVENTS

The duration of prograde metamorphism in the English River Subprovince can currently only be bracketed for the eastern Lac Seul area. Granulite-grade metamorphism developed there during a period of approximately 32 my. This range is defined by an upper age limit of 2692 Ma, the age of pre-D_1 orthopyroxene-bearing leucosomes in metasedimentary migmatite, and a lower age limit of 2660±20 Ma, the time of emplacement of late-orogenic granite-granodiorite during the waning stages of static granulite-grade metamorphism (Krogh et al. 1976).

GEOPHYSICS

Regional and local scale geophysical features show the distinctiveness of the English River Subprovince relative to the neighbouring subprovinces. This section briefly reviews

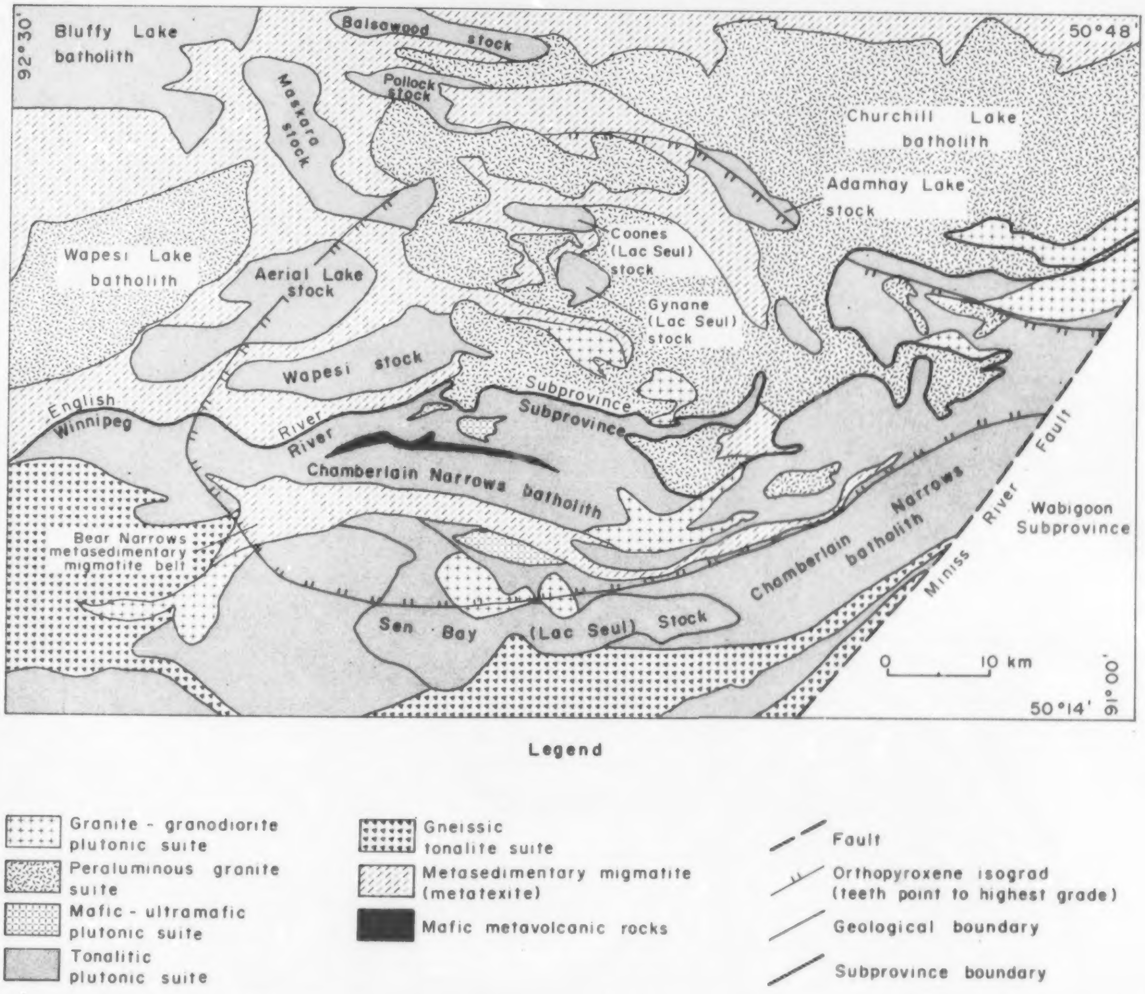
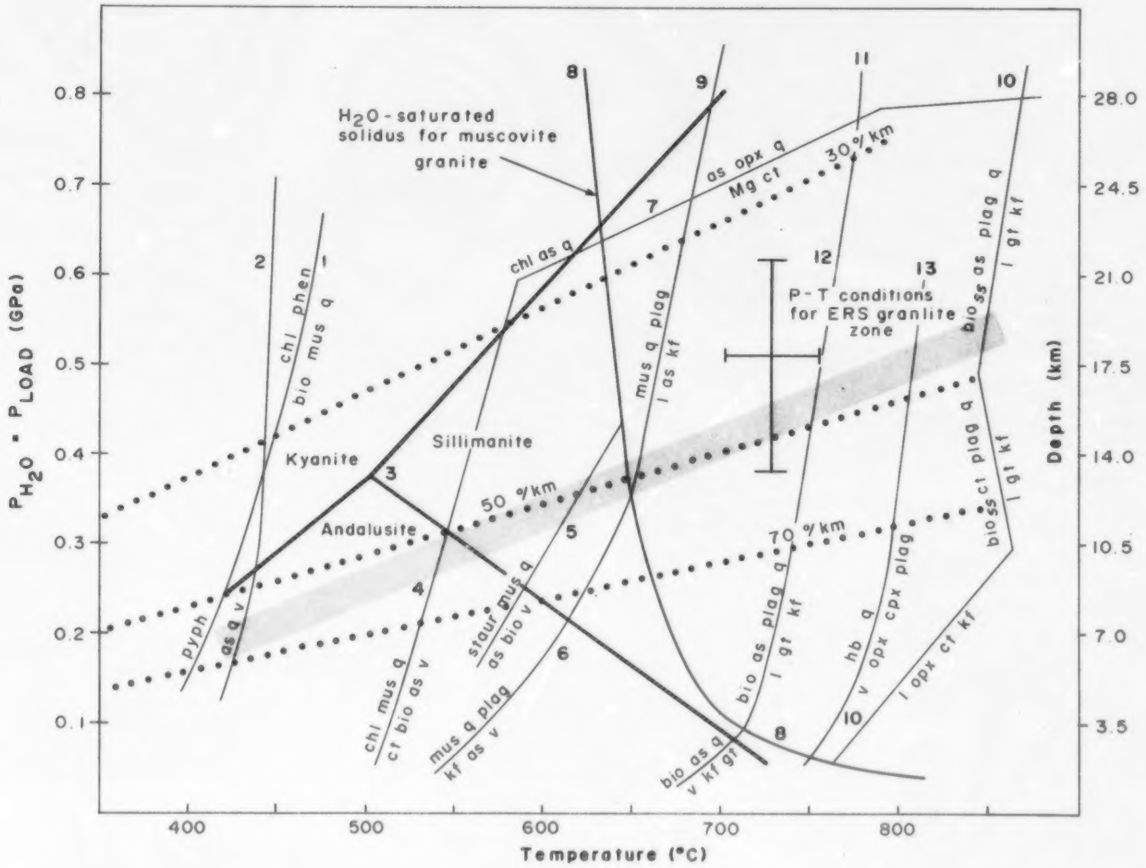

92°30'
Bluffy Lake batholith
Balsawood stock
Maskara stock
Pollock stock
50°48'
Churchill Lake batholith
Adamhay Lake stock
Coones (Lac Seul) stock
Wapesi Lake batholith
Aerial Lake stock
Gynane (Lac Seul) stock
Wapesi stock
English River Subprovince
Winnipeg River Subprovince
Chamberlain Narrows batholith
Bear Narrows metasedimentary migmatite belt
Sen Bay (Lac Seul) Stock
Minniss River Fault
Wabigoon Subprovince
0 10 km
50°14'
91°00'
Legend
Granite - granodiorite plutonic suite
Peraluminous granite suite
Mafic - ultramafic plutonic suite
Tonalitic plutonic suite
Gneissic tonalite suite
Metasedimentary migmatite (metatexite)
Mafic metavolcanic rocks
Fault
Orthopyroxene isograd (teeth point to highest grade)
Geological boundary
Subprovince boundary

Figure 7.9. Detailed geology of the boundary region of the English River and Winnipeg River subprovinces, showing location of granulite-grade isograd in the eastern Lac Seul granulite zone.



Legend

▨ Path of region metamorphism

••• Metamorphic isograd

Abbreviations

as	aluminosilicates
bio	biotite
chl	chlorite
cpx	clinopyroxene
ct	cordierite
ERS	English River Subprovince
gt	garnet
hb	hornblende
kf	potassium feldspar
l	liquid
Mg	Magnesium
mus	muscovite
opx	orthopyroxene
phen	phengite
plag	plagioclase
pyph	pyrophyllite
q	quartz
ss	solid solution
staur	staurolite
v	vapour phase

1	Hemley (1967)
2	Brown (1971)
3	Holdaway (1971)
4	Hirschberg and Winkler (1968)
5	Richardson (1970)
6	Thompson (1976)
7	Newton et al. (1974)
8	Huang and Wyllie (1973)
9	Kerrick (1972)
10	Vielzeuf and Holloway (1988)
11	Clemens and Wall (1981)
12	Perkins and Chipera (1985); Chipera and Perkins (1988)
13	Lower temperature boundary for granulite - grade metamorphism based upon dehydration of common hornblende in the presence of quartz (Binns 1969) and extrapolated from 3 to 6 kb.

Figure 7.10. Pressure/GPa (depth/km)-temperature/ °C grid on which is plotted: metamorphic reaction and melting curves, mineral stability fields, geothermal gradients, and an inferred regional metamorphic gradient now preserved as a southward increase in conditions of regional metamorphism from the Uchi Subprovince into the English River Subprovince.

gravitational, seismic, magnetic and radiometric data collected during 30 years of geophysical investigation in the English River Subprovince.

Gravity

One of the foremost geophysical features is a regionally prominent positive gravity anomaly which is coincident with the length of the subprovince for about 450 km (Innes 1960; *see* maps 2592 to 2595, map case). This anomaly, which is up to 20 mgal above the regional datum, is interrupted by 2 narrow areas of northeast-striking gravity lows. These lows are coincident with the Miniss River Fault and the Wapesi Lake batholith and another gravity low (cause unknown), that is north-trending and approximately coincident with an axis between Attwood and Marshall lakes (*see* Figure 7.2). This area lies to the east of the Whitewater–Mojikit lakes granulite zones, is characterized by an area 80 by 230 km of tonalitic rocks distributed along the English River Subprovince–Uchi Subprovince boundary east of Atwood Lake. The combination of these tonalites, extensive peraluminous rocks (e.g., Abamasagi batholith, *see* Figure 7.2) and a locally preserved greenstone belt (Melchett Lake), coupled with an absence of granulite-grade metamorphism, might explain the negative gravity anomaly in this area. At the extreme eastern end of the English River Subprovince, lying underneath the Paleozoic cover rocks, there exists a broad oval gravity high which is spatially associated with the confluence of the subprovincial gravity high and which is coincident with the north-northeast-striking Kapuskasing Structural Zone. Towards the western end of the English River Subprovince, the regional positive gravity anomaly bifurcates at the Ontario–Manitoba boundary part of which becomes a southwest-striking arm. The significance of the bifurcation is unknown, because it is apparently not related to lithologic patterns at surface.

The regional gravity pattern contrasts with that of the geologically similar Quetico Subprovince (*see* maps 2593 and 2594, map case; Cheadle 1982; Percival 1989; Williams, this volume). The contrast may reflect, in part, differing depths to the Mohorovicic discontinuity beneath the Quetico Subprovince (Young et al. 1986, 35 to 42 km) and the English River Subprovince (Hall and Hajnal 1973, 30 to 36 km). Detailed gravity work (Gupta and Barlow 1984; Gupta and Wadge 1986), was undertaken west of long. 94°W in the English River Subprovince and in the adjacent Uchi Subprovince and Winnipeg River Subprovince. A transect across the English River Subprovince along the Red Lake Road revealed a residual positive anomaly of 15 mgal, centred over the English River Subprovince and a modelled thickness of 10 km for the metasedimentary prism. In contrast to the English River Subprovince, the adjacent Winnipeg River Subprovince is characterized by much lower gravity values. A regional gravity survey conducted by Gupta and Wadge (1986) included the northern part of the English River Subprovince between 91°45′W and 94°30′W, where a steep residual gravity gradient of 12 to 15 mgal was determined to coincide with the Sydney Lake–Lake St. Joseph boundary fault. These authors found also that large tonalite-trondhjemite-granodiorite suite plutons, such as the Bluffy Lake batholith, exhibit little gravity response relative to the host metasedimentary rocks; they modelled a sheet thickness of only 1.5 to 3 km.

In some respects, the anomaly pattern along the English River Subprovince–Uchi Subprovince boundary resembles those marking fundamental structural boundaries in the Archean Limpopo mobile belt in southern Africa (Barton and Key 1981). There, steep gravity gradients have been interpreted to result from the thrusting of dense granulite-grade lower crustal rocks upon lower density upper crust (Barton and Key 1981, p.192–194). Thomas (1989) suggested that the positive-negative anomaly pair along the English River Subprovince–Uchi Subprovince boundary may, in fact, represent a relict collisional suture zone.

The positive regional gravity anomaly over much of the English River Subprovince relates to a number of features:

1. some combination of the occurrence of dense, granulite-grade rocks (MacKidd 1973; Urquhart 1976; Thurston and Breaks 1978; Chipera 1985)
2. thinner total crust relative to the adjacent Uchi Subprovince or Winnipeg River Subprovince and upwarping of the Mohorovicic discontinuity (Hall 1971, 1974; Hall and Hajnal 1973; Gupta and Wadge 1986)
3. zones of mesocratic and melanocratic restite remaining at depth after extraction, and emplacement of leucocratic S-type peraluminous granite magma at higher crustal levels, now removed by erosion

Seismic Investigations

The crustal configuration of the crust beneath the English River Subprovince and adjacent subprovinces has been evaluated by the seismic reflection and refraction surveys of Hall (1968, 1971) and Hall and Hajnal (1969, 1973). These works defined a peculiar crustal structure, which, based upon a 2 layer seismic model, features an upward concavity in the Riel or Conrad discontinuity and a spatially coincident downward convexity in the Mohorovicic discontinuity. The trace of these curves lies within the English River Subprovince (Beakhouse 1977) and was later confirmed by Hall and Brisbin (1982). Upper crustal thickness is modelled at 21 to 26 km, whereas the Mohorovicic discontinuity varies in depth from 30 to 36 km. Reinterpretation of the seismic data used in a 3 layer crustal model (Hall and Brisbin 1982), involved the addition of a 6 km thick intermediate layer having a P-wave seismic velocity of 6.9 km/s. This layer might represent a concentration of restite unaffected by extraction and emplacement of felsic peraluminous granite magma to a higher crustal level. Whereas, the upper crustal layer is known to be composed of metasedimentary migmatite and tonalitic plutonic rocks, the lithologic composition of the lower crustal layer is equivocal. Christensen and Fountain (1975) have shown that the P- and S-wave velocities of Hall and Hajnal (1973) are consistent with granulite-grade rocks, or with rocks of mafic to intermediate igneous composition.

Seismic tomographic data (Grand 1987) indicate that the western part of the Superior Province has a 400 km thick lithospheric root, that has a high shear wave velocity. An analysis of shear wave anisotropy by Silver and Chan (1988) indicates a strong east-trending anisotropy associated with the English River Subprovince–Uchi Subprovince boundary region which may have been recorded within both crustal and mantle rocks during the Kenoran Orogeny. Further application of these types of seismic research may indicate more precisely, the age of thickness variation and fabric within the mantle lithosphere which probably result from crustal accretionary processes.

Magnetic Data

The aeromagnetic lineament maps of Wilson (1971) and Gower (1978) reveal that the English River Subprovince is dominated by a strong linear to locally curvilinear magnetic grain (*see* Maps 2584 to 2587, map case). This grain parallels the regional foliation and contrasts, both with the more irregular magnetic trends of greenstone-granite subprovinces, and the higher magnetic intensity pattern featured by more broadly spaced anomalies in the Winnipeg River Subprovince.

A clear distinction between the English River Subprovince and Winnipeg River Subprovince may be made with frequency-filtered data which is between 60 and 4000 km in wavelength (Hall 1968, 1974). The English River Subprovince has a characteristic, generally flat magnetic field of low regional intensity which correlates with the low magnetic susceptibility of metasedimentary migmatite (Hall et al. 1979, average 0.1 A/m). The large peraluminous granite plutons (e.g., Wapesi Lake batholith) are also of low magnetic intensity, which renders them virtually indistinguishable from the enveloping metasedimentary migmatite.

Exceptions to the strongly anisotropic, yet homogeneous magnetic pattern of the English River Subprovince are local, moderately magnetic areas with clusters of linear and oval positive anomalies. These are associated with banded iron formation units of considerable continuity, with zones of granulite-grade metamorphism, and with larger plutons of the tonalite-trondhjemite-granodiorite suite and massive granodiorite to granite suite, such as the Bluffy Lake, Rowdy Lake and Fletcher Lake batholiths (*see* Figure 7.2). Urquhart (1976) and Urquhart and West (1979) examined an area of moderate magnetic intensity in which the total field pattern exhibited numerous, narrow linear anomalies up to 30 km in length and with a maximum amplitude of 700 nT above base level. Urquhart (1976) demonstrated that the high surface magnetic field strength is closely correlated with areas of granulite-grade metamorphism and that the magnetic anomalies are coincident with numerous narrow units of orthopyroxene-bearing amphibolite. The flat-lying attitude of these thin units, coupled with the macroscopic fold closures magnetically delineated by Urquhart (1976) favoured the interpretation that recumbent folds are present in the region.

Radiometric Data

A considerable part of the English River Subprovince has been surveyed for potassium, uranium and thorium (OGS–GSC 1977). Large masses of peraluminous granite, such as the Wapesi Lake batholith, exhibit a strong spatial correlation with contour maxima for potassium and total counts. Metasedimentary migmatite, in contrast, has a relatively flat radiometric response.

TECTONO-MAGMATIC SYNTHESIS

The English River Subprovince, as well as the geologically similar Quetico Subprovince, constitute a type of high-grade terrain that is rare or absent in most Archean cratons (Williams 1990). This section will review the geologic development of the English River Subprovince and incorporate it into a tectonic model. Many geologic events influenced development of the English River Subprovince and its boundary zones; these are summarized in the time-space diagram for the English River Subprovince and boundary zones involving the Winnipeg River Subprovince, Uchi Subprovince and Wabigoon Subprovince (*see* Maps 2575 to 2583, map case).

Previous Tectonic Models

In the last 20 years, many tectonic models have addressed crustal development in the Superior Province. These, as summarized by Blackburn (1980) and Card (1990), fall into 2 major categories:

1. "fixist" models which treat supracrustal and intrusive rocks as being essentially autochthonous stratigraphic sequences subjected to vertical tectonics (e.g., Wilson et al. 1974; Goodwin 1977; Young 1978; Henke 1984; Percival and Sullivan 1988)
2. "mobilist" models which consider the configuration of greenstone belts and enveloping granitoids to be chiefly the product of horizontal, plate-driven tectonics, brought into juxtaposition via accretion, subduction and collision of terrane components (e.g., Douglas and Price 1972; Talbot 1973; Langford and Morin 1976; Burke et al. 1976; Blackburn 1980; Hoffman 1989; Percival and Williams 1989; Williams 1990)

Fixist, or nonplate tectonic models envisage the English River Subprovince to be a sedimentary basin which is developed upon sialic crust and is positioned between 2 volcanic arcs, now represented by the Uchi Subprovince and Wabigoon Subprovince. Vertical subsidence led to thickening and metamorphism of the clastic pile which ultimately resulted in migmatization. This model cannot account for the thrust interleaving of supracrustal and older tonalitic rocks distributed along the English River Subprovince–Winnipeg River Subprovince and English River Subprovince–Uchi Subprovince boundary zones, nor the wide variation in absolute ages of such volcanic assemblages (*see* Beakhouse, this volume and Stott and Corfu, this volume).

Mobilist models have gained credence in the last decade. Structural data, geochronology and volcano-sedimentary facies relationships have favoured uniformitarian, mobilist tectonic models (e.g., Kay and Stott 1985; Davis et al. 1988; Devaney and Williams 1989; Williams 1990). Langford and Morin (1976) suggested by analogy with the Cordillera of western North America, that the striped pattern of Archean subprovinces was produced by accretion of island arc and intervening sedimentary basins on to a sialic craton now represented by the Sachigo and Berens River subprovinces (*see* Thurston, Osmani et al., this volume). Blackburn (1980) and Blackburn et al. (1985) proposed a plate tectonic model for the Wabigoon Subprovince which envisaged development of oceanic crust by the rifting apart of sialic crust.

Chipera (1985) modified the Langford and Morin (1976) model by suggesting that the Winnipeg River Subprovince represents an allochthonous terrane which was partly thrust over the southern part of the English River Subprovince. This inference has not been fully tested against field, geochronologic, or metamorphic data. More recently, Beakhouse (1988) and Beakhouse and McNutt (1991) proposed that a magmatic arc-continent collision zone (Sioux Lookout domain), at least 250 km in strike length, developed along the Winnipeg River Subprovince–Wabigoon Subprovince boundary between 2700 and 2710 Ma. This zone is marked by thrust stacking of supracrustal rocks (Beakhouse 1988; Davis et al. 1988; Davis 1990; Blackburn et al., this volume) and has several similarities with the English River Subprovince, including high-grade metasedimentary migmatite hosting a linear belt of S-type peraluminous granite plutons and cogenetic rare-element pegmatites (Breaks 1989, 1990).

Current Tectonic Model

The tectonic model proposed below explains many of the tectonic and magmatic events and is a modification of the plate tectonic hypothesis for part of the northwestern Superior Province developed by Blackburn (1980) and Blackburn et al. (1985). It departs from earlier models by invoking an older sialic microcontinent between the English River Subprovince and Wabigoon Subprovince.

Langford and Morin (1976) interpreted the English River Subprovince to be an 800 km long fore-arc basin to an island-arc system represented by the Uchi Subprovince. The metasedimentary belt may represent an accretionary prism (Hoffman 1989); however, the supportive evidence of the type used in a similar interpretation of the Quetico Subprovince (Devaney and Williams 1989; Williams 1990) is not forthcoming.

The following sections discuss the processes of sedimentation, magmatism and metamorphism pertaining to the geologic development of the English River Subprovince.

SEDIMENT SOURCE

The provenance of the epiclastic sedimentary rocks deposited in the English River (fore-arc?) basin was possibly a combination of metavolcanic and tonalitic material; this is inferred from lithologic and stratigraphic-facing data along the English River Subprovince–Uchi Subprovince boundary (Westerman 1977; Breaks et al. 1978; Meyn and Palonen 1980; Kay and Stott 1985; Breaks and Bond, in prep.). Stratigraphic relations indicate that the low- to medium-grade English River Subprovince metasedimentary rocks unconformably overlie 2714^{+5}_{-2} million-year-old (Corfu and Stott 1989) Cycle III mafic to felsic metavolcanic rocks in the western part of Lake St. Joseph (Meyn and Palonen 1980; Kay and Stott 1985) and stratigraphically overlie 2738^{+5}_{-2} million-year-old Cycle III metavolcanic rocks in the Slate Lake area (Nunes and Thurston 1980; Breaks and Bond, in prep.). The English River Subprovince basin may have developed before the deposition of these volcanic sequences, but sedimentation cannot have started before 2760 Ma, the age of the oldest detrital zircon (Krogh et al. 1976). Whereas, Breaks et al. (1978) interpreted the English River Subprovince metasedimentary rocks to be composed of a mixture of felsic to intermediate plutonic and mafic-ultramafic metavolcanic detritus, van de Kamp and Beakhouse (1979) and Beakhouse (1985) considered that felsic volcanic centres were a major source of detritus. All of these lithic groups probably acted as source material as inferred from their presence in clast populations of boundary zone conglomerates (Breaks and Bond 1976).

TIMING OF SEDIMENTATION

Sedimentation occurred between 2760 and 2698 Ma; this overlaps the depositional age ranges of clastic rocks situated to the south of the English River Subprovince (2714 to 2706 Ma) in the Warclub metasedimentary rocks of the Wabigoon Subprovince (Davis 1990) and is slightly earlier than the Quetico metasedimentary rocks (2702 to 2687 Ma) (Percival 1989; Davis et al. 1990).

TIMING OF MAGMATISM

The initial phases of metasedimentary belt plutonism occurred in the English River Subprovince from 2.65 to 2.7 Ga and produced regionally extensive tonalite sheets and more localized syntectonic plutons of the massive granodiorite to granite suite of probable I-type genesis (White and Chappell 1977). Tonalites emplaced during this time interval have no apparent equivalents in the Winnipeg River Subprovince; Beakhouse and McNutt (1991) have indicated that tonalitic rocks are consistently older (2.83 to 3.17 Ga) than most in the English River Subprovince metasedimentary migmatite mass.

SOURCES AND CAUSES OF MAGMATISM

The tonalites in the English River Subprovince are regarded as products of the partial melting of tholeiitic oceanic crust. Such protolith material maybe similar to that preserved as the 2745 million-year-old Separation Lake greenstone belt and undated, compositionally similar amphibolitic mafic metavolcanic units exposed in the eastern Lac Seul region. Tonalitic magmatism could have been generated by sub-duction and partial melting of oceanic crust along a north-dipping zone. This zone would have been positioned beneath recently deposited and imbricated English River Subprovince sedimentary rocks, into which the bodies of tonalite would have been emplaced . Such an orientation of the subduction zone would have favoured the more or less vertical rise of relatively young (2.65 to 2.698 Ga) tonalite-trondhjemite-granodiorite suite plutons into the

English River Subprovince metasedimentary migmatite mass.

Magmas of the syntectonic massive granodiorite to granite suite may be derived from the partial melting of tectonically thickened crust as was proposed for the generation of type-4 plutons in the Winnipeg River Subprovince by Beakhouse et al. (1989); the composition of the source is as yet undetermined.

SUBPROVINCE BOUNDARIES

The northern boundary of the English River Subprovince against the Uchi Subprovince is faulted. Originally it may have been a zone of lithologic transition between an arc-like environment to the north, and a sedimentary basin to the south. In a modern plate tectonics scenario, such a broad distinction between 2 adjacent environments is often a complex one modified by thrusting and wrenching, the telescoping of an original variation in facies and the juxtaposing of rocks of disparate origins. The combination of geophysical, structural and metamorphic contrasts across the boundary leads the author to suspect that it is of fundamental tectonic significance.

The southern boundary of the English River Subprovince, against the Wabigoon Subprovince and the Winnipeg River Subprovince is much less well documented, and in places may also represent a sedimentary basin-volcanic arc transition. Against the Winnipeg River Subprovince, in the west, the Bird River–Separation Lake belt marks the boundary zone, which is interpreted to be a zone of shearing. In the east, the English River Subprovince is bounded by the Wabigoon Subprovince; in places this is an abrupt, faulted contact, elsewhere, it is transitional and poorly documented.

TECTONICS

In the present model invoking north-dipping subduction, the arrival from the south of a buoyant *circa* 3 billion-year-old sialic microcontinent, presently forming part of the Winnipeg River Subprovince, at the trench zone would effectively arrest subduction prior to the initial 2692 million-year-old anatexis of metasedimentary rocks in the English River Subprovince. The juxtaposition of the Winnipeg River Subprovince microcontinent with the south edge of the English River Subprovince might have been the cause for the regional compression and imbricate stacking of thrust sheets along the English River Subprovince–Winnipeg River Subprovince boundary. This process of stacking may be demonstrated in the eastern Lac Seul boundary zone area, where it has led to the tectonic intercalation of *circa* 3 billion-year-old allochthonous tonalites with much younger (2760 to 2698 Ma) metasedimentary rocks. The allochthonous nature of the older (3 Ga) tonalites at Chamberlain Narrows and Sen Bay, relative to the associated metasedimentary rocks is permitted by a lack of greater than 3 billion-year-old zircons within those sedimentary rocks. This interpretation is supported by the work of Sanborn-Barrie (1988), who concluded that Winnipeg River Subprovince tonalitic rocks were obliquely transported over the English River Subprovince metasedimentary rocks during the D_2 compressional event.

CAUSES OF ENGLISH RIVER SUBPROVINCE METAMORPHISM

Many recently proposed models have been devised to explain the cause of metamorphism in high-grade metasedimentary gneiss belts, which are mainly Phanerozoic and Proterozoic in age (e.g., Wickham and Oxburgh 1985; Lux et al. 1986; Gordon, in press). The following mechanisms have been invoked as a means of generating high-temperature, low-pressure regional metamorphic assemblages: crustal underplating by basaltic or tonalitic magmas (Wells 1980; Wickham and Oxburgh 1987); crustal thickening due to overstacking or understacking of thrust sheets (England and Thompson 1984; Davy and Gillet 1986; Zen 1988); emplacement of granite magmas (Lux et al. 1986; England and Thompson 1986); and crustal extension and attendant rise of isotherms (McKenzie 1978; Thompson 1981; Wickham and Oxburgh 1985).

Mafic magma has been considered by several workers (Wells 1980; Wickham and Oxburgh 1987) as an important source of conductive heat essential to produce low-pressure, high-temperature metamorphism. Accretion of mafic magma to the base of the crust is considered by Thompson (1981) as "the most efficient means of simultaneously heating and evolving tectonically thickened crust so as to expose andalusite-sillimanite or kyanite-sillimanite facies series metamorphic conditions at surface." This thermal-controlling influence has been dismissed by several workers for many metamorphic terranes for reasons such as this: the absence of mafic plutonic rocks that slightly predate or are coeval with the peraluminous granites (in contrast to the model of Foster and Hyndman 1990). Although the English River Subprovince contains a few mafic plutons, this mechanism may be supported by the interpreted upward convexity of the Mohorovicic discontinuity being a result of increased accretion of mafic magmas to the base of the English River Subprovince crust relative to adjacent subprovinces.

In recent years, evidence of horizontal tectonism has steadily been gathered in the northwestern Superior Province (Westerman 1977; Thurston and Breaks 1978; Poulsen et al. 1980; Davis et al. 1988; Beakhouse et al. 1989; Williams 1990). Within the Winnipeg River Subprovince, close to the English River Subprovince, Westerman (1977) demonstrated that tonalitic gneisses, 3170 Ma (Corfu 1988) in age, and now isolated in the structural domes of the Cedar–Clay lakes area, were emplaced over metasedimentary migmatite. The metasedimentary rocks may also have been tectonically transported, from the English River Subprovince, during D_2 deformation and prior to development of the domes and emplacement of 2758 million-year-old Cliff Lake potassium feldspar megacrystic granodiorite (Westerman 1977; Beakhouse 1983; Corfu 1988).

A potential method of inducing high-grade metamorphism and migmatization along the English River Subprovince–Winnipeg River Subprovince boundary zone would be by increasing crustal thickness by thrusting. If thickening involved emplacement of relatively hot, deep level tonalitic slabs onto cooler metasedimentary rocks (Wells 1980), the occurrence of the highest grades close to the English River Subprovince–Winnipeg River Subprovince tectonic boundary (*see* Figure 7.9) would be explained.

Chipera (1985) explained the distribution and intensity of metamorphism in terms of a 400°C thermal perturbation with a slight rise in pressure (0.4 to 0.6 GPa) relative to adjacent greenstone-rich subprovinces. He concluded that emplacement of magmas at 900°C coupled with a 5 km uplift of the English River Subprovince relative to adjacent greenstone-rich subprovince was necessary.

Tonalitic suite plutons emplaced at 2.65 to 2.7 Ga represent the only granitic rocks regionally distributed in the English River Subprovince that could have supplied sufficient heat to effect high-grade metamorphism. Field evidence, however, indicates that many of these rocks have been regionally metamorphosed and locally show evidence of partial melting. Furthermore, the regional orthopyroxene isograd cuts across tonalite-trondhjemite-granodiorite suite plutons in several areas (*see* Figure 7.9). The only members of the tonalite-trondhjemite-granodiorite suite that can definitely be linked to granulite-grade metamorphism are the narrow dikes associated with volumetrically insignificant static granulite-grade selvages.

Gosnold and Perkins (1989), on the basis of two-dimensional finite difference modelling, concluded that intrusion of mafic peraluminous granite magma was capable of producing large areas of granulite-grade metamorphism in the English River Subprovince within a remarkably brief time span of 2 my. Geochronologic data, however, indicates that such metamorphism developed over a much longer interval of approximately 32 my which makes such a model unrealistic. The one-dimensional thermal modelling of anatexis made by Zen (1988) in thrust-thickened sialic crust predicts that anatexis typically lags behind the causative tectonic event by 20 to 60 my. This compares with the English River Subprovince which recorded a protracted and episodic process of leucosome generation from 2692 Ma, until after 2668 Ma.

Furthermore, the mafic peraluminous granites are considered by Breaks et al. (1978) and Breaks and Bond (in prep.) to represent the product of a higher temperature, more extensive partial melting of metasedimentary rocks, but these granites are greatly subordinate to their leucocratic and hololeucocratic peraluminous granite counterparts. The peraluminous granite suite, which has a predominantly minimum melt composition, has been previously demonstrated in this review to be a "consequence" of the dynamic phase of granulite-grade metamorphism and not a "cause" of it.

Recent experimental and numerical modelling work (LeBreton and Thompson 1988; Vielzeuf and Holloway 1988; Patino–Douce et al. 1990) indicates that the amount of anatectic peraluminous granite melt deemed necessary to exceed the critical melt fraction (Arzi 1978; van der Molen and Patterson 1979) can, in the absence of an externally derived fluid, be generated only from a metasedimentary protolith by means of undersaturated partial melting induced by dehydration of biotite at temperatures of 850° to 900°C. Such temperatures have not, to date, been recorded in any geothermometric work done in the English River Subprovince. Therefore, it may be that the large peraluminous granite masses associated with some granulite nodes (e.g., Churchill Lake batholith, *see* Figure 7.2) were either derived at a deeper crustal level in the metasedimentary prism or that "wet" anatexis (fluid-present melting) effected the generation of large volumes of anatectic melt in the geothermometrically determined temperature interval of 650° to 750°C of Chipera and Perkins (1988). However, as pointed out by Clemens and Vielzeuf (1987, p.3299) "if metamorphism was fluid present with a high H_2O, then all lithologic types with quartz and feldspars should show signs of extensive melting, once temperature reached 650°C". Such temperatures and rock types are widespread in the English River Subprovince (Breaks et al. 1978; Henke 1984; Chipera 1985) yet most metasedimentary migmatite is of the low leucosome fraction type (protometatexite and metatexite). Therefore, anatexis is probably mainly of the fluid-absent type.

The larger English River Subprovince granulite-grade zones bear no clear spatial relationship to that of the peraluminous granite or tonalitic suite plutons contained within such metamorphic zones (*see* Figure 7.9). If one of these intrusive suites was the prime source of conductive heat required to produce high-grade metamorphism in the fashion of Lux et al. (1986), it follows that a clear positive correlation would be expected.

Crustal extension may have played an as yet undefined role in the development of the English River Subprovince sedimentary basin and its subsequent metamorphism. This process has been suggested by Thompson (1981) as a rapid means of inducing higher temperatures and a rise in isotherms. Structural evidence supporting crustal extension in the English River Subprovince is minimal and is relegated to the presence of north-striking pregranulite mafic dikes within tonalitic pluton segments of the English River Subprovince–Winnipeg River Subprovince boundary zone. Such dikes may represent Archean diabase whose emplacement was conceivably induced by a tensional regime. Furthermore, crustal extension does not adequately explain the widespread presence of low-pressure metamorphism in the Superior Province, which is characterized by such metamorphism following crustal shortening.

Thermal relaxation of the differentially tectonically thickened crustal pile led to temperatures of 650° to 750°C, that after cooling and erosion are presently manifested as a thermal anticline. Uncertainty, however, exists as to whether these temperatures represent peak values (Chipera 1985, p.99). Peak metamorphic conditions may not have been preserved within the English River Subprovince granulite-grade zones, a common situation for granulites on a global basis (Frost and Chacko 1989). Thermal relaxation produced high-grade metamorphism and episodic migmatization, for at least 32 million years, and outlasted collision-related deformation as shown by the overprinting of the eastern Lac Seul granulite zone upon the English River Subprovince–Winnipeg River Subprovince boundary.

Archean igneous activity in the English River Subprovince was terminated by the intrusion of pegmatites of the massive granodiorite to granite suite at about 2650±40 Ma (Krogh et al. 1976). These rocks were derived by partial melting of older tonalitic crustal components (Beakhouse 1983).

CONCLUSIONS

The major conclusions of this chapter are:

1. The English River Subprovince displays many of the sedimentologic, structural, magmatic and metamorphic properties of fore-arc accretionary prisms.
2. The sedimentary prism developed over a period of less than 20 my, and was derived from a volcanic-dominated supply emanating from the north, within the Uchi Subprovince.
3. Structural and magmatic-thickening processes induced widespread high-grade metamorphism.
4. The contact between the Uchi Subprovince and the English River Subprovince is a fundamental boundary, the character of which is reflected in the structural, metamorphic and geophysical contrasts across it.

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Chapter 8

Winnipeg River Subprovince

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Abstract

The Winnipeg River Subprovince is a dominantly granitoid domain within the central part of the western Superior Province. The Subprovince is flanked to the south by a granite-greenstone domain (western Wabigoon Subprovince) and to the north by a metasedimentary migmatite domain (English River Subprovince) and a greenstone domain (Bird River Subprovince).

Highly metamorphosed and deformed supracrustal rocks composed of massive and pillowed basalts and minor cherty and ferruginous chemical sedimentary rocks are intruded by, and tectonically included in, tonalitic gneisses and plutons that range in age from 2.83 to 3.17 Ga. These supracrustal remnants, though not extensive and only fragmentarily preserved, record the earliest stage in the development of the Winnipeg River Subprovince and are among the oldest rocks yet recognized in the Superior Province in Ontario. Metaplutonic rocks that are dominantly of tonalitic composition are subdivided into a gneissic suite and a tonalitic plutonic suite. The gneissic suite is very heterogeneous, comprising tonalitic phases containing abundant mafic enclaves and being intruded by variously deformed granitic/pegmatitic sheets. The tonalitic plutonic suite is more homogeneous than the gneissic suite and, in most areas, forms distinct plutons, although it is temporarily and genetically related to the tonalite component of the gneissic suite. Tonalitic components of both suites are interpreted to have originated by the partial melting of tholeiitic basalt at lower crustal or upper mantle depths. Granodioritic to granitic plutons, the granitic suite, and minor dioritic to granodioritic plutons, the mafic suite, were emplaced between 2.66 and 2.71 Ga. The granitic suite is interpreted to have originated by the partial melting of the older metatonalites in tectonically thickened crust.

The transition from tonalitic to granitic plutonism follows a period of geological quiescence in the Winnipeg River Subprovince which ended with the development of large-scale thrusts and/or nappes at 2707 ± 8 Ma. This tectonic thickening is interpreted to be responsible for subsequent doming caused by gravitational instability and the initiation of high-grade metamorphism and intracrustal melting (granitic suite), which was followed by protracted cooling of the crust.

INTRODUCTION

The Winnipeg River Subprovince is a granitoid terrane located in the southwestern part of the Superior Province. The Subprovince was originally regarded as part of the English River Subprovince as defined by Wilson and Brisbin (1964). Wilson (1971) regarded the English River "block" to consist predominantly of amphibolite to granulite-facies metasedimentary rocks and anatectic granitoid rocks. Subsequent studies demonstrated that the English River Subprovince consists of 2 distinct terranes (Beakhouse 1977; Breaks et al. 1978) for which separate subprovince status was proposed (Hall and Brisbin 1982). The most widely used terminology today is that of Card and Ciesielski (1986) who preserve the terms English River Subprovince to refer to the northern, dominantly sedimentary terrane and refer to the southern, dominantly granitoid terrane as the Winnipeg River Subprovince. These authors also proposed subprovince status for the Bird River Subprovince which occurs at the Winnipeg River–English River subprovince boundary in southeastern Manitoba and can be traced as a narrow, continuous volcanic septum for at least 70 km into Ontario.

As presently defined and exposed, the Winnipeg River Subprovince comprises approximately 15 000 km². It is bounded to the north by the English River Subprovince and to the south by the western Wabigoon Subprovince (Figure 8.1). To the west, the Winnipeg River Subprovince is overlain by Paleozoic strata of the Williston Basin and extends an unknown distance beneath these strata (*see* Hall 1974; McCabe et al. 1990). To the east, the Winnipeg River Subprovince apparently pinches out in the vicinity of Savant Lake; however, it is possible that the Winnipeg River Subprovince and the dominantly granitoid central domain of the Wabigoon Subprovince (Blackburn et al., this volume) which share certain of the same characteristics (Beakhouse 1988a and references cited therein) may be related.

DESCRIPTION OF LITHOLOGICAL UNITS

The generalized distribution of rock types is presented in Figure 8.2 and on the accompanying geological map of Ontario (*see* Maps 2541 to 2545 inclusive, map case). Most rocks within the Winnipeg River Subprovince can be broadly grouped into 5 suites (*see* Maps 2541 to 2545 inclusive *in* map case for the unit numbers in brackets) namely: 1) supracrustal rocks (predominantly units 5 and 7); 2) gneissic tonalite suite (unit 11); 3) foliated tonalitic suite (unit 12); 4) granitic plutonic suite (unit 15); and 5) diorite-monzonite-granodiorite suite (unit 14). Generalized summaries of the nature of these suites, with special emphasis on field and relative age relationships, petrography, geochemistry and origin are given in the following sections.

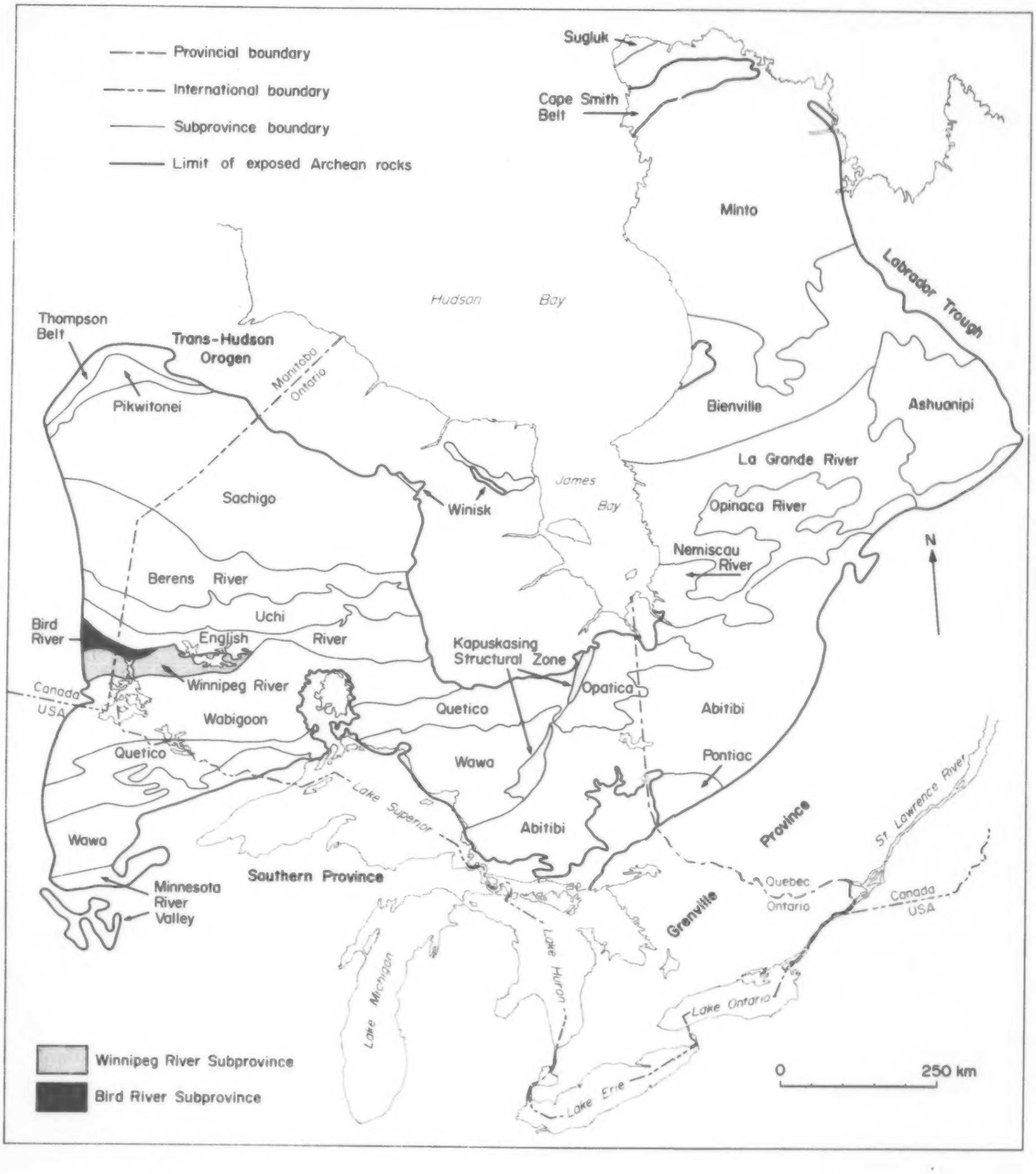

Provincial boundary
International boundary
Subprovince boundary
Limit of exposed Archean rocks
Sugluk
Cape Smith Belt
Minto
Labrador Trough
Hudson Bay
Thompson Belt
Trans-Hudson Orogen
Manitoba
Ontario
Pikwitonei
Bienville
Ashuanipi
La Grande River
James Bay
Sachigo
Winisk
Opinaca River
Nemiscau River
N
Berens River
Uchi
Bird River
English River
Kapuskasing Structural Zone
Winnipeg River
Opatica
Canada
USA
Wabigoon
Quetico
Quetico
Abitibi
Wawa
Pontiac
Lake Superior
Wawa
Abitibi
Province
St Lawrence River
Southern Province
Minnesota River Valley
Quebec
Ontario
Canada
USA
Grenville
Lake Huron
Lake Michigan
Lake Ontario
Lake Erie
Winnipeg River Subprovince
Bird River Subprovince
0
250 km

Figure 8.1. Geological map of the Superior Province illustrating the location and regional setting of the Winnipeg River Subprovince (*modified from* Card and Ciesielski 1986).

Supracrustal Rocks

Supracrustal rocks can be further subdivided into disrupted and migmatized remnants within the gneissic suite and those rocks which form more coherent stratigraphic units. As elaborated below, these 2 modes of occurrence of supracrustal rocks have contrasting field relationships and ages. Consequently these rocks are interpreted to have different tectonic significance and are discussed separately.

SUPRACRUSTAL REMNANTS

Supracrustal remnants within the gneissic suite are not abundant, but they are widely, though heterogeneously, distributed. Although the remnants exhibit a range of abundances on regional, local and outcrop scales, these remnants are estimated to represent approximately 5 to 10% of the volume of the gneissic sequences that contain them, and less than 2% of the Winnipeg River Subprovince as a whole. The field relations, which are discussed below, suggest that these remnants are among the oldest supracrustal rocks in the Superior Province. These remnants are commonly small and disrupted, but can be up to 1 km thick and 5 km long. Some of these units, such as those near the Dalles pluton (Gower 1978) and in the Ross Lake area, are thin, (20 to 40 m) but are laterally continuous for tens of kilometres.

Mafic metavolcanic rocks predominate within the supracrustal remnants. All primary textures have been obliterated and these rocks have been transformed into amphibolites (plagioclase + hornblende ± quartz ± clinopyroxene) or mafic granulites (plagioclase + orthopyroxene + clinopyroxene + hornblende ± quartz) (Beakhouse et al. 1983). In some areas characterized by the relatively lower grade assemblage, rare clinopyroxene develops in response to compositional controls most commonly proximal to calcic pods and the more mafic selvages of pillows. In other areas, such as the Daniels tonalite, clinopyroxene is more widely developed in mafic supracrustal remnants, and presumably reflects more extreme conditions of metamorphism. Garnet is rarely present and may in part be related to the breakdown of hornblende accompanying partial melting (Gower 1978). Biotite is a minor, though widespread, retrograde alteration product of hornblende and pyroxene. Biotite is conspicuously developed along amphibolite-granitoid contacts and is interpreted to be due to infiltration of potassium-bearing hydrous fluids during the emplacement of late granitic rocks. Other retrograde minerals recognized include fibrous amphibole, epidote, white mica and chlorite. Accessory minerals include sulphides, magnetite, sphene, apatite and zircon.

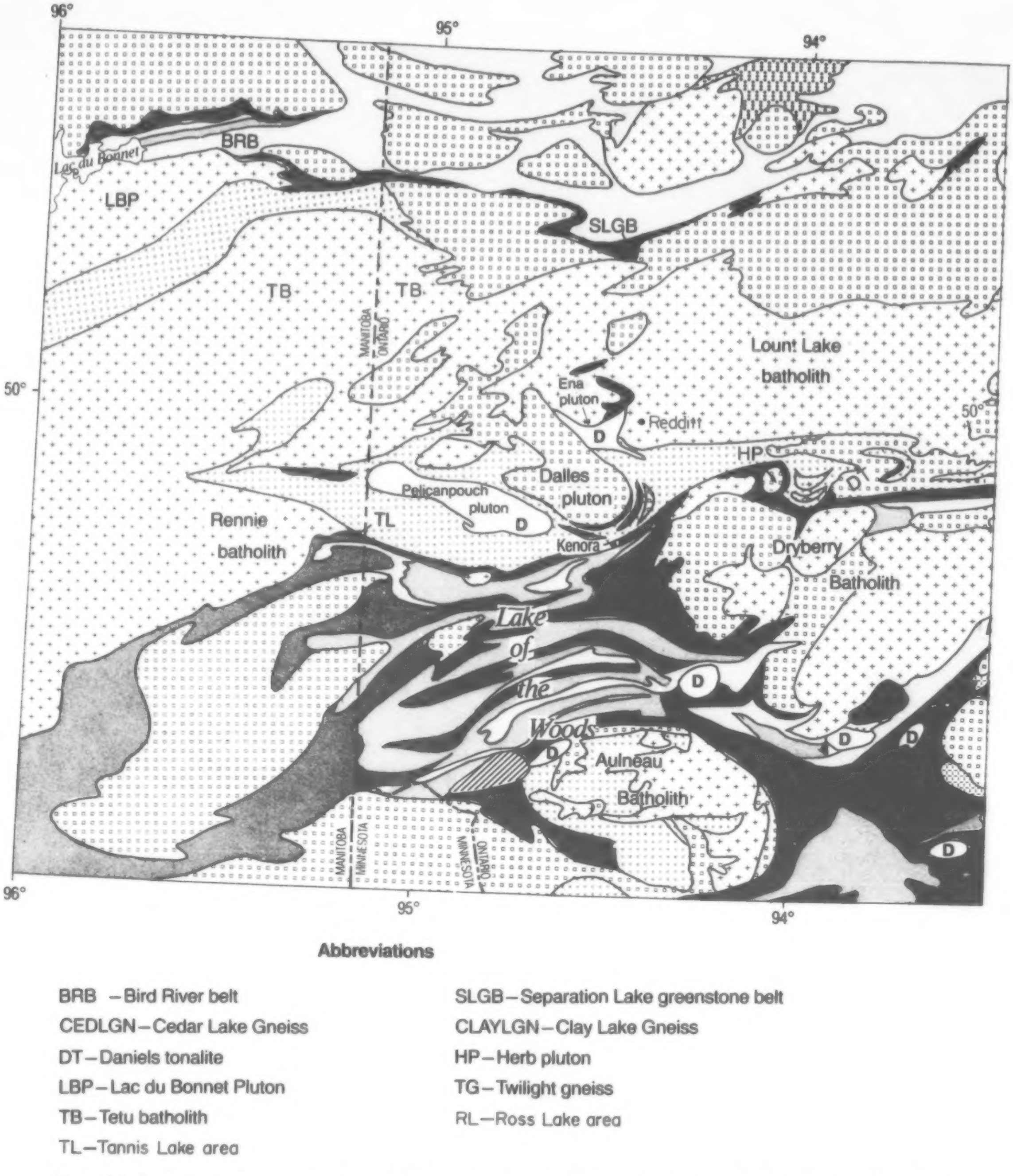
Primary structures are difficult to identify because of intense deformation, disruption by younger intrusive rocks and local partial melting. Locally, structures such as pillows are readily recognizable even at the highest metamorphic grades (Figure 8.3a). With increasing intensity of deformation, pillows become progressively flattened and elongated, and the rock is ultimately transformed into a banded amphibolite with thin (less than 1 cm), variously disrupted, more mafic lenses and layers representing selvage material (Figure 8.3b). Calcic pods, which in the Kenora area consist of clinopyroxene + quartz + plagioclase ± grossularite/andradite garnet (Gower 1978), are found within recognizable pillow lavas and banded amphibolite, and persist to advanced stages of anatexis. Similar pods occur locally within lower grade volcanic sequences in the Superior Province and represent metamorphosed calcareous interpillow material and/or diagenetic introduction of calcitic alteration into gas cavities or scoria. These pillowed and banded amphibolites are interlayered with compositionally and texturally similar massive amphibolite and medium- to coarse-grained gabbro, and represent massive and pillowed basaltic flows with related gabbro sills. Confidence in this interpretation is locally enhanced by the presence of thin chert-magnetite iron formation units (Figure 8.3c) within these mafic units.

Compositionally, mafic metavolcanic rocks from the Winnipeg River Subprovince are subalkalic tholeiitic basalts (Westerman 1977; Gower 1978; Beakhouse, unpublished data). Although there is some evidence for the addition of alkali elements, the compositions of these basalts are closely comparable to lower metamorphic grade Archean basaltic rocks from elsewhere in the Superior Province and modern basaltic rocks. Gower (1978) compared abundances of the relatively immobile elements and concluded that the mafic volcanic remnants in the Kenora area closely resemble modern island arc tholeiites and ocean floor basalts.

Metavolcanic rocks, having other than a mafic composition, are either rare or are difficult to recognize due to their highly deformed and metamorphosed condition. Rocks of ultramafic composition occur within the supracrustal remnants, but in the absence of textural criteria indicative of a mode of origin, they could be either komatiitic metavolcanic rocks, cumulates derived from mafic magmas, or the residue from partial melting of previously less mafic rocks. Since one might expect that, given the common preservation of primary structures in tholeiitic basalts, komatiites would be equally well preserved, it is concluded that komatiites are either not present or are much less abundant than tholeiitic basalt in the supracrustal remnants of the Winnipeg River Subprovince. Westerman (1977) reported 2 analyses of metavolcanic rocks comparable to basaltic komatiites, but these are from a large roof pendant in a granitic pluton and cannot be unequivocally related to the supracrustal remnants. Gower (1978) recognized "fragmental" rocks of intermediate composition from several outcrops in the Kenora area, but pointed out that the fragmental nature is not clearly indicative of a supracrustal origin.

Metasedimentary rocks are a very minor component (less than 5%) of the supracrustal remnants. Chert-magnetite iron formation and chert are the most abundant and readily identifiable metasedimentary components. Although extensively recrystallized, these rocks preserve delicate, thin bedding (*see* Figure 8.3c).

The recognition of clastic metasedimentary rocks is more problematic because primary sedimentary structures are difficult to identify. Some of the deformational/metamorphic banding superficially resembles bedding, and the intensity of the processes that lead to the development has so severely transposed any original bedding and modified the


96°
95°
94°
50°
50°
BRB
Lac du Bonnet
LBP
SLGB
TB
TB
MANITOBA
ONTARIO
Lount Lake batholith
Ena pluton
Redditt
D
HP
Dalles pluton
Pelicanpouch pluton
TL
Rennie batholith
Kenora
Dryberry Batholith
Lake of the Woods
Aulneau Batholith
MANITOBA
MINNESOTA
ONTARIO
MINNESOTA
96°
95°
94°

Abbreviations

BRB –Bird River belt
CEDLGN–Cedar Lake Gneiss
DT–Daniels tonalite
LBP–Lac du Bonnet Pluton
TB–Tetu batholith
TL–Tannis Lake area
SLGB–Separation Lake greenstone belt
CLAYLGN–Clay Lake Gneiss
HP–Herb pluton
TG–Twilight gneiss
RL–Ross Lake area

Figure 8.2. Generalized geological map of the Winnipeg River Subprovince indicating the locations and units discussed in the text.

shape and size of grains that any primary structures that may have existed have been obscured. Detailed examination of relatively well-preserved supracrustal remnants indicates clastic metasedimentary rocks are less abundant than chemical metasedimentary rocks.

Probable clastic metasedimentary rocks have been recognized primarily on the basis of their compositions. Most notable are rocks characterized by aluminous mineral assemblages with greater than 30% biotite ± muscovite ± garnet ± sillimanite ± chlorite. Such compositions may be indicative of a pelitic sedimentary origin (wacke or shale); but in the absence of the preservation of primary sedimentary structures, other origins are also conceivable (e.g., metamorphosed alteration of volcanic rocks or residue from vapour-present partial melting). Gower (1978) reported 2 examples of meta-arkose (consisting of up to 66% quartz

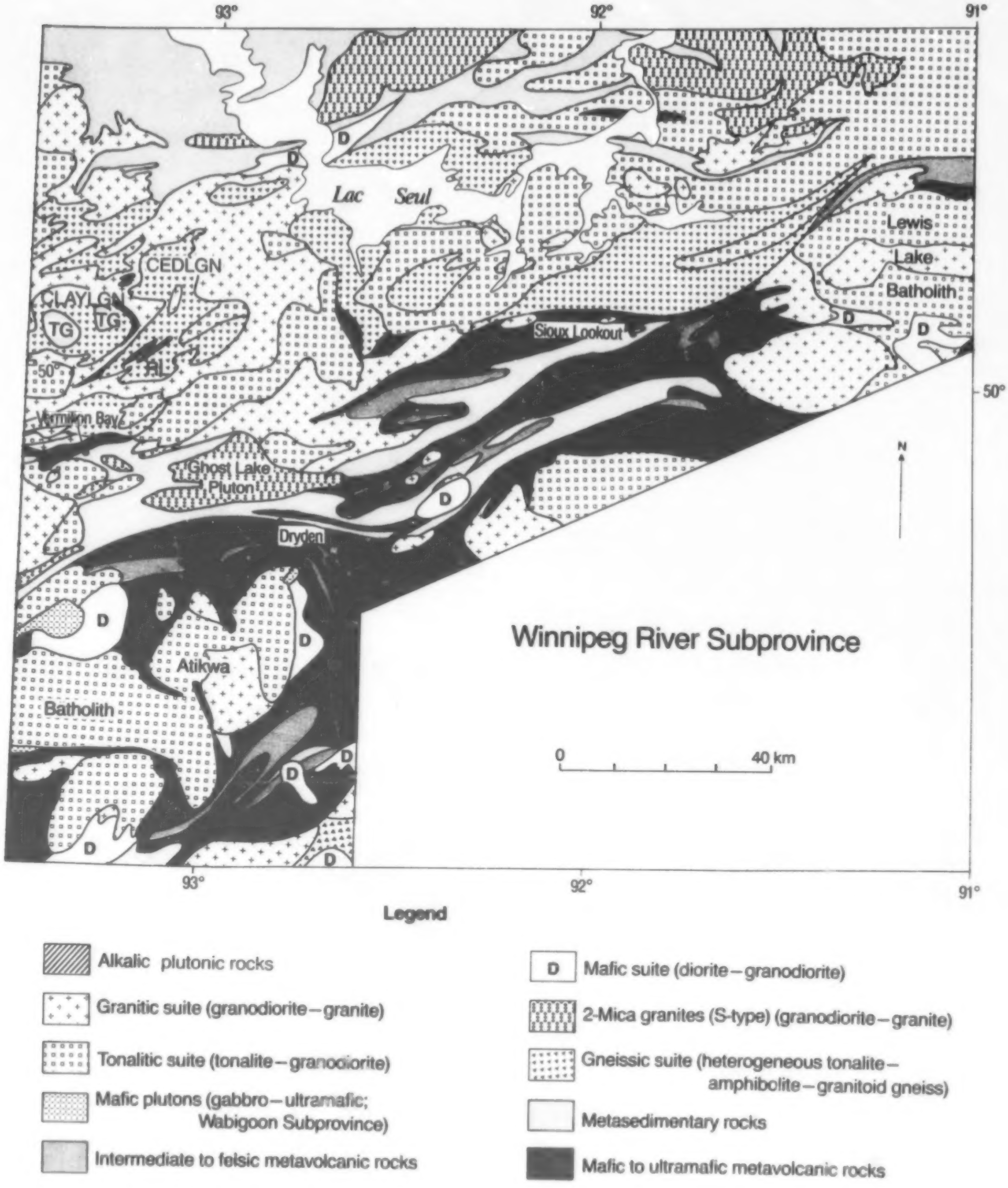

93°
92°
91°
Lac Seul
CEDLGN
CLAYLGN
TG
TG
RL
50°
Vermilion Bay
Ghost Lake Pluton
Dryden
Sioux Lookout
Lewis Lake Batholith
D
Atikwa Batholith
Winnipeg River Subprovince
0
40 km
N
50°
Legend
Alkalic plutonic rocks
Granitic suite (granodiorite–granite)
Tonalitic suite (tonalite–granodiorite)
Mafic plutons (gabbro–ultramafic; Wabigoon Subprovince)
Intermediate to felsic metavolcanic rocks
Mafic suite (diorite–granodiorite)
2-Mica granites (S-type) (granodiorite–granite)
Gneissic suite (heterogeneous tonalite–amphibolite–granitoid gneiss)
Metasedimentary rocks
Mafic to ultramafic metavolcanic rocks

Figure 8.2. Generalized geological map of the Winnipeg River Subprovince indicating the locations and units discussed in the text.

accompanied by plagioclase and biotite); one associated with a biotite + plagioclase + garnet rock, has isotopically heavy oxygen ($\delta^{18}O = 9.4$, Longstaffe and Gower 1983), providing further evidence for a sedimentary protolith. However, such rocks are rare, and in most cases it is difficult to completely discount the possibility that quartz-rich compositions could reflect premetamorphic introduction of secondary silica.

The age of the supracrustal remnant suite has not been determined directly. Many remnants are tectonic inclusions in metaplutonic rocks and even relative ages with respect to the rocks, that contain them are commonly obscured. Nevertheless, the remnants are interpreted to be the oldest rocks within the subprovince since they are intruded by the oldest metaplutonic rocks (Westerman 1977; Gower 1978; Beakhouse 1983). The U-Pb zircon ages of these metaplutonic rocks span a broad range (2.83 to 3.17 Ga, *see* below). Supracrustal remnants are most abundant in the younger of these sequences, but are present to some degree in all of them. For example, in the Kenora area, supracrustal enclaves constitute approximately 15% of the 2.83 to 2.88 Ga gneissic rocks, whereas these enclaves are less abundant in the pre-3.0 Ga Cedar Lake Gneisses. The significance of this observation is unclear; it may simply be an artifact of the limited geochronological data base or may indicate that the older metaplutonic rocks contain fewer supracrustal enclaves because: 1) they have been more thoroughly disrupted; 2) younger plutonic rocks were preferentially emplaced in zones with abundant supracrustal remnants; or 3) supracrustal rocks were less abundant during the time of emplacement of the older plutonic complexes. The latter possibility highlights the unresolved problem of whether, in spite of being the oldest rock in any particular area, some of the supracrustal enclaves might be younger than the oldest plutonic rocks (i.e., there may be 1 old, prior to 3.17 Ga, supracrustal sequence, or several sequences older than 2.83 Ga, one of which is greater than 3.17 Ga).

Although the supracrustal remnants comprise a minor component of the subprovince, they may represent the oldest crust in the entire Superior Province. Preservation is fragmentary, and perhaps biased, but some generalizations are warranted. The presence of pillowed basalt and thinly bedded chemical sedimentary rocks indicates a submarine depositional environment. These early supracrustal rocks are overwhelmingly of tholeiitic basaltic composition and closely compare in chemistry to many later Archean basalts and modern tholeiitic basalts from various tectonic environments. There is broad agreement that modern tholeiitic basalts originate by moderate degrees of partial melting at shallow depths in a peridotitic mantle (Green and Ringwood 1967); similar interpretations have been presented for better preserved Archean sequences (Hart et al. 1970) and a uniformitarian interpretation is preferred for the mafic supracrustal remnants. Unequivocal clastic metasedimentary and intermediate to felsic metavolcanic rocks are rare, suggesting that there was little or no reworking of older sialic crust. However, as discussed below, there is evidence to suggest that rocks, similar in composition to the mafic supracrustal remnants, may have been melted at lower crustal to shallow mantle depths to produce the tonalitic plutonic rocks which intrude the remnants.

SUPRACRUSTAL BELTS

In addition to the supracrustal remnants discussed above, there are several coherent metavolcanic and metasedimentary sequences within and marginal to the Winnipeg River Subprovince. Examples include: the Bird River–Separation Lake belt, the Twilight gneiss of the Cedar Lake area, and the Zealand metasedimentary rocks. Mafic metavolcanic rafts and roof pendants in young plutonic complexes (e.g., at Vermilion Lake, Redditt area, and Cliff Lake) are not discussed.

The Bird River–Separation Lake belt occurs along the Winnipeg River–English River subprovince boundary and has been afforded subprovince status (Card and Ciesielski 1986). Within Manitoba, this belt extends for 60 km, is up to 8 km wide, and displays considerable lithologic diversity. Within Ontario, the belt consists predominantly of massive and pillowed mafic metavolcanic rocks (Figure 8.3d) together with minor cherty and ferruginous chemical metasedimentary rocks forming a thin, laterally persistent septum. The septum is less than 200 m thick for much of its 70 km length in Ontario, except in the Separation Lake area where the belt achieves a thickness of several kilometres. To the north, the Separation Lake part of the septum is in fault contact with metasedimentary rocks of the English River Subprovince to the north. The southern contact with the Winnipeg River Subprovince is characterized by intrusive, migmatitic contacts with younger plutonic rocks along much of its length. At Greer Lake in Manitoba, the belt is in fault contact with tonalitic gneiss (Cerny et al. 1981).

Geological relationships in Manitoba are critical in deciphering the relationship of the Bird River–Separation Lake belt to the English River Subprovince and Winnipeg River Subprovince. In Manitoba, mafic to felsic metavolcanic rocks alternate with a variety of clastic metasedimentary rocks in a series of thin laterally persistent, largely fault-bounded belts (Cerny et al. 1981). The laterally persistent mafic septum in Ontario is continuous with the Lamprey Falls Formation of the Bird River belt in Manitoba. Some of the metasedimentary rocks of the Bird River belt have been correlated with the migmatitic metasedimentary rocks of the English River Subprovince (Cerny et al. 1981). Most of the sedimentation is closely related to volcanism; although, significantly, one of the stratigraphically highest formations, the Flanders Lake Formation contains abundant plutonic detritus (Cerny et al. 1981). Constraints on the age of volcanism and sedimentation in the Bird River–Separation Lake belt are confined to the Bird River area where Timmins et al. (1985) report U-Pb ages of 2741 Ma for the Peterson Creek Formation, 2745±5 Ma for the Bird River sill which represents a minimum age for the Lamprey Falls and Eaglenest Lake formations (assuming correlations in Manitoba are valid) and 2715 Ma for a diorite emplaced into the Bernic Lake Formation. Based on this data, together with the correlation of parts of the Bird River belt with the metasedimentary migmatite of the English River Subprovince, these rocks are interpreted to be significantly younger than the supracrustal remnants in the Winnipeg River Subprovince, and, closely comparable in age to other supracrustal sequences which are marginal to the Winnipeg River Subprovince.

This interpretation requires either that the Bird River–Separation Lake belt unconformably overlies the Winnipeg River Subprovince or that the former is allochthonous. The locally faulted contact of the belt, absence of evidence for an unconformity and volcaniclastic character of the basal Eaglenest Lake Formation suggest that the latter is the preferred interpretation. The addition of a metaplutonic detrital

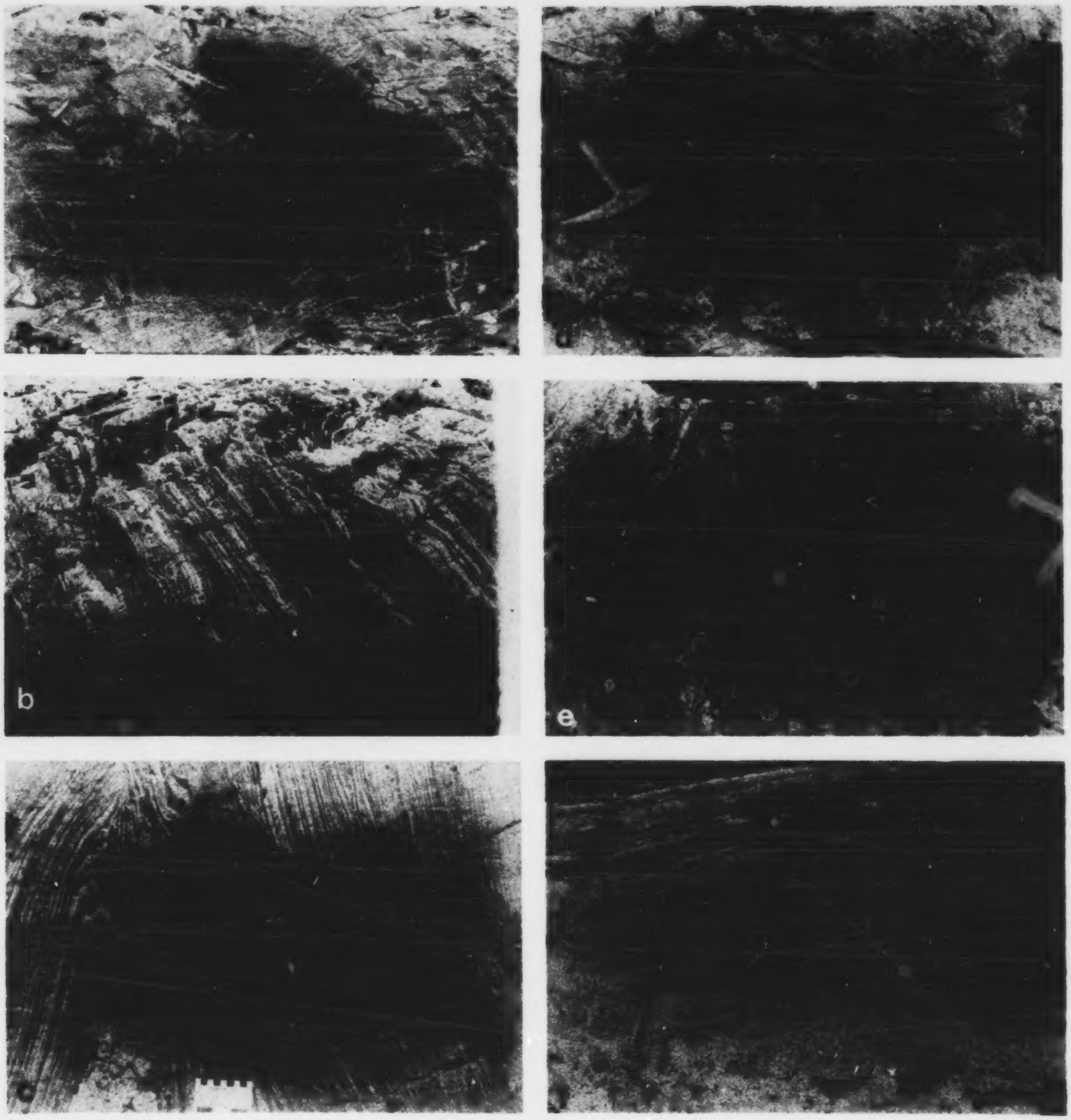


Figure 8.3. Outcrop photographs of supracrustal rocks from the Winnipeg River Subprovince: **a)** moderately deformed, pillowed basalt, Cedar Lake area; **b)** strongly deformed pillowed basalt with strongly transposed and attenuated selvages and calc-silicate clots and cut by leucotonalite sheets, Kenora area; **c)** folded chert-magnetite iron formation from the Trail Lake series (Westerman 1977) in the Cedar Lake area; **d)** pillowed basalt, Separation Lake greenstone belt; **e)** incipiently migmatized, interbedded wacke-siltstone (this example, from the Pakwash Lake area of the English River Subprovince, is similar to that occurring in the Twilight gneiss); and **f)** migmatitic wacke-siltstone, Twilight gneiss.

component, near the top of the sequence, suggests that early development of the belt occurred in a volcanic-dominated environment away from the influence of older sialic crust but that sedimentation continued during tectonic transport and deposition of the uppermost part of the stratigraphy records the interaction with a metaplutonic terrane such as that exposed in the Winnipeg River Subprovince.

Card and Ciesielski (1986) have suggested that the Bird River–Separation Lake belt should be elevated to subprovince status. Based on precedence of subdivision of the Superior Province elsewhere, this appears justified. For example, relationships between the Bird River belt and the English River Subprovince are analogous to those between the Beardmore–Geraldton belt of the Wabigoon Subprovince and the Quetico Subprovince (Blackburn et al., this volume). In the latter case, the Beardmore–Geraldton belt is excluded from the Quetico Subprovince in spite of the fact that sedimentary units now in the Beardmore–Geraldton belt and Quetico Subprovince were originally part of a continuous depositional tract (Devaney and Williams 1989). Similarly, relationships between the Bird River belt and the Winnipeg River Subprovince are analogous to those between the Winnipeg River Subprovince and the Sioux Lookout domain; this domain is considered to be part of the Wabigoon Subprovince (Beakhouse 1988b). Consequently, the elevation to subprovince status proposed by Card and Ciesielski (1986) is valid by default since the application of principles used elsewhere in the Superior Province excludes the Bird River belt from both adjacent subprovinces. However, this conclusion does not mean that adjacent subprovinces are not related in several ways. The continuity of metasedimentary units between the Bird River and English River subprovinces demonstrates that for at least a part of their development, they were a continuous depositional tract. Although the early histories of the Bird River and Winnipeg River subprovinces are apparently independent, late sedimentation (postdating or synchronous with tectonic emplacement) (Flanders Lake Formation) and granitic plutonism may be manifestations of a shared history. These problems point to the inherent limitations of the subprovincial subdivision of the Superior Province.

Migmatitic, pelitic metasedimentary rocks (Zealand metasedimentary rocks) near Dryden have been interpreted to be part of the Winnipeg River Subprovince based largely on their high metamorphic grade (Breaks et al. 1978). These rocks are correlated stratigraphically with lower greenschist facies metasedimentary rocks in the Sioux Lookout and northern Lake of the Woods areas (Blackburn et al., this volume) and are therefore regarded herein to be part of the Wabigoon Subprovince. These rocks have been interpreted to be part of an allochthonous terrane (Davis et al. 1988; Beakhouse 1988b) that is analogous in certain respects to the Bird River Subprovince.

The only relatively large areas of undisrupted supracrustal rocks not intruded by gneissic tonalite that occur within the Winnipeg River Subprovince, are those metasedimentary rocks exposed in the Clay Lake area. These rocks, referred to as the Twilight gneiss by Westerman (1977), occur in the core of domal structures and are structurally overlain by metaplutonic rocks. The Twilight gneiss is similar to migmatitic metasedimentary rocks from the Pakwash Lake area of the English River Subprovince (van de Kamp and Beakhouse 1979). Sedimentary structures, principally bedding defined by variations in mineralogy and texture (Figure 8.3e) and, rarely, mineralogical grading interpreted to reflect original gradations in clay mineral content, indicate that this unit has a sedimentary origin. Isotopically "light" oxygen ($\delta^{18}O = 7.3$ to 9.2) is unlike that of other metasedimentary rocks in the Superior Province and has been interpreted to reflect isotopic exchange with adjacent orthogneisses during the high-grade metamorphism (Longstaffe and Schwarcz 1977). The relatively uniform thickness (5 to 20 cm) and even, continuous nature of the original sedimentary beds, together with the inferred graded bedding, are consistent with deposition from turbidity currents. The Twilight gneiss is composed principally of plagioclase, quartz and biotite with minor garnet and orthopyroxene in addition to a wide variety of accessory minerals (Westerman 1977). The aluminous composition (biotite = 10 to 33%; garnet = 0 to 10%) suggests that these rocks were originally wackes and shales.

Widespread, thin, commonly concordant leucocratic lenses with melanocratic margins are interpreted to be *in situ* partial melts of the metasedimentary rocks (Figure 8.3f). The presence of garnet and orthopyroxene, together with the evidence that these rocks have undergone partial melting, indicates that lowermost granulite-facies metamorphic conditions were attained. Westerman (1977) estimates that peak metamorphic conditions were in the range 650° to 750°C and 450 to 700 MPa.

Westerman (1977) noted the similarity in rock type and structural history between the Twilight gneiss and paragneisses of the English River Subprovince and proposed that they may in fact be correlative. However, these rocks are also very similar to migmatitic paragneisses in the Dryden area, and in either case, there is no conclusive evidence on which to base such a correlation.

Despite uncertainty in correlation, both the English River Subprovince paragneisses and Zealand metasedimentary rocks have been interpreted to be younger than the gneissic granitoid rocks of the Winnipeg River Subprovince (Beakhouse 1985). The possibility that a relatively young age might apply to the Twilight gneiss, which structurally underlies older (since interpreted to be 3.0 to 3.2 Ga; Beakhouse 1983; Corfu 1988) metaplutonic rocks, led Westerman (1977) to conclude that metaplutonic basement rocks were tectonically interleaved with younger metasedimentary rocks. The abrupt nature of the contact, with no paragneiss inclusions in the metaplutonic rocks and no metaplutonic rocks in the paragneiss, is consistent with such an interpretation.

The arguments discussed above suggest that the larger, more coherent metavolcanic and metasedimentary sequences within and marginal to the Winnipeg River Subprovince postdate the emplacement of the metaplutonic rocks. All of these sequences can, with varying degrees of confidence, be interpreted to be allochthonous. The metaplutonic rocks apparently did not contribute detritus to the

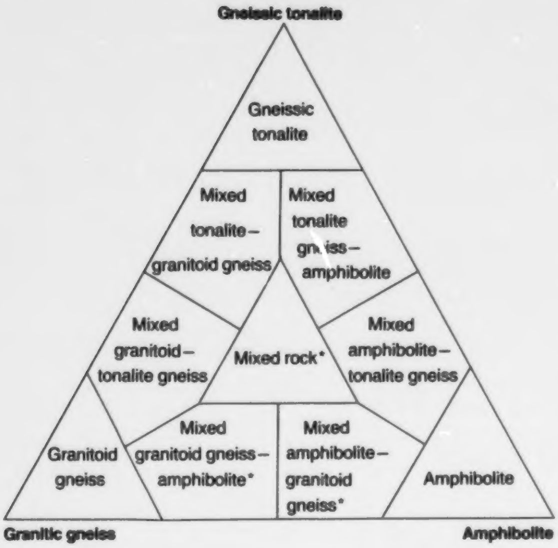

Gneissic tonalite
Gneissic tonalite
Mixed tonalite–granitoid gneiss
Mixed tonalite gneiss–amphibolite
Mixed granitoid–tonalite gneiss
Mixed rock*
Mixed amphibolite–tonalite gneiss
Granitoid gneiss
Mixed granitoid gneiss–amphibolite*
Mixed amphibolite–granitoid gneiss*
Amphibolite
Granitic gneiss
Amphibolite

Figure 8.4. Ternary diagram illustrating the nomenclature used to describe the gneissic suite. The characteristics of each of the end member components are discussed in the text. Fields designated by an asterisk are less common (*modified from* Gower 1978).

stratigraphically lower parts of these sequences; however, the upper parts of some of these sequences contain plutonic clasts which indicate derivation from an older granitoid terrane (interpreted to be the Winnipeg River Subprovince). On this basis, the timing of this tectonic transport is interpreted to postdate the intrusion of tonalitic (meta)plutonic rocks (approximately 2.83 to 3.17 Ga; *see* following discussion) and predate the emplacement of granitic plutons (2.66 to 2.71 Ga) which intrude them.

Gneissic Suite

The gneissic suite is a compositionally and texturally heterogeneous assemblage of moderately to strongly deformed metaplutonic rocks with subordinate supracrustal enclaves. A ternary classification scheme (Figure 8.4, *modified from* Gower 1978) forms the basis of this description of these highly variable assemblages.

Tonalitic (*sensu lato*) metaplutonic rocks are the most abundant component of the gneissic suite. These rocks have, in order of decreasing relative abundance, the mineralogical composition of tonalite, leucotonalite and quartz diorite (IUGS classification, Streckeisen 1976; Figure 8.5). These rocks are fine- to medium-grained and are usually equigranular. These rocks are characterized by foliation and/or lineation in addition to a weakly to strongly developed gneissosity. Gneissosity is defined mainly by lenses and layers of coarser grained leucotonalite as well as mafic schlieren (Figures 8.6a, 8.6b). Many of the mafic schlieren are disrupted enclaves. Others are associated with the margins of the coarser grained leucotonalite lenses and layers and the pair represent complementary products of partial melting or metamorphic differentiation of their tonalitic host (*see* Figure 8.6b). The layered character of these rocks is often enhanced by the development of mylonitic banding and emplacement of intrusive sheets, both of which are commonly parallel or subparallel to the gneissosity. Multiple generations of gneissic fabric development are common (Figures 8.6c, 8.6d). The mineralogical and chemical composition of these rocks are discussed under tonalitic plutonic rocks.

The amphibolite component of the gneissic suite includes rocks of basaltic composition, regardless of their origin. Some amphibolites are demonstrably metavolcanic, rocks (described above) and others are disrupted mafic dikes, but in most cases the origin is indeterminate. However, amphibolites that are interpreted to be of metavolcanic origin are more common and widespread than those that clearly are dikes, and it is probable that a high percentage of those of indeterminate origin are metavolcanic rocks. Although some of the metavolcanic enclaves are distinctive by virtue of their heterogeneity and features such as calc-silicate clots, most of the mafic metavolcanic rocks and mafic dikes, as well as the amphibolites of indeterminate origin, have similar mineralogy and chemical composition. Greater than 90% by volume of these rocks are plagioclase (approximately 25 to 45%) and mafic minerals. The mafic mineralogy varies with metamorphic grade; typical assemblages comprise hornblende, hornblende + clinopyroxene or clinopyroxene + orthopyroxene + hornblende. Biotite and, less commonly, chlorite occur as retrograde alteration products of these mafic minerals, especially along contacts with granitoid rocks. Minor amounts of quartz (less than 8%) are almost always present. Trace accessory minerals include opaque minerals, sphene and zircon .

The third apex of the ternary classification scheme (*see* Figure 8.4) is granitoid gneiss, whereas Gower (1978) used the term granitic pegmatoid gneiss in deference to pegmatitic textures which characterize these phases in the Kenora area. On a more regional basis, although pegmatitic textures are common, rocks having a more uniform, equigranular, fine- to medium-grained texture are also present. These rocks are typically leucocratic (colour index less than 5) with common, heterogeneously distributed mafic schlieren. Local crosscutting relationships indicate that they postdate some of the gneissosity development, but are typically strongly foliated to gneissic. Granitoid gneiss includes a wide variety of rocks that have diverse origins; some examples are discussed below.

1. Medium-grained, equigranular leucotonalite bands and lenses occur within amphibolite and tonalite. Some have mafic selvages and geometries which suggests an origin by *in situ* partial melting of the enclosing rocks (*see* Figure 8.6b; Gower 1978; Beakhouse 1983).
2. Leucogranodiorite and leucogranite sheets range from medium-grained, equigranular to coarsely pegmatitic (Gower 1978), and occur mainly within gneissic tonalite. These rocks are typically concordant or subconcordant with respect to the gneissosity. In some cases, these rocks have mafic selvages and geometries that suggest they result from *in situ* melting of the tonalites (Westerman 1977), but in other cases they are clearly intrusive into the tonalites.

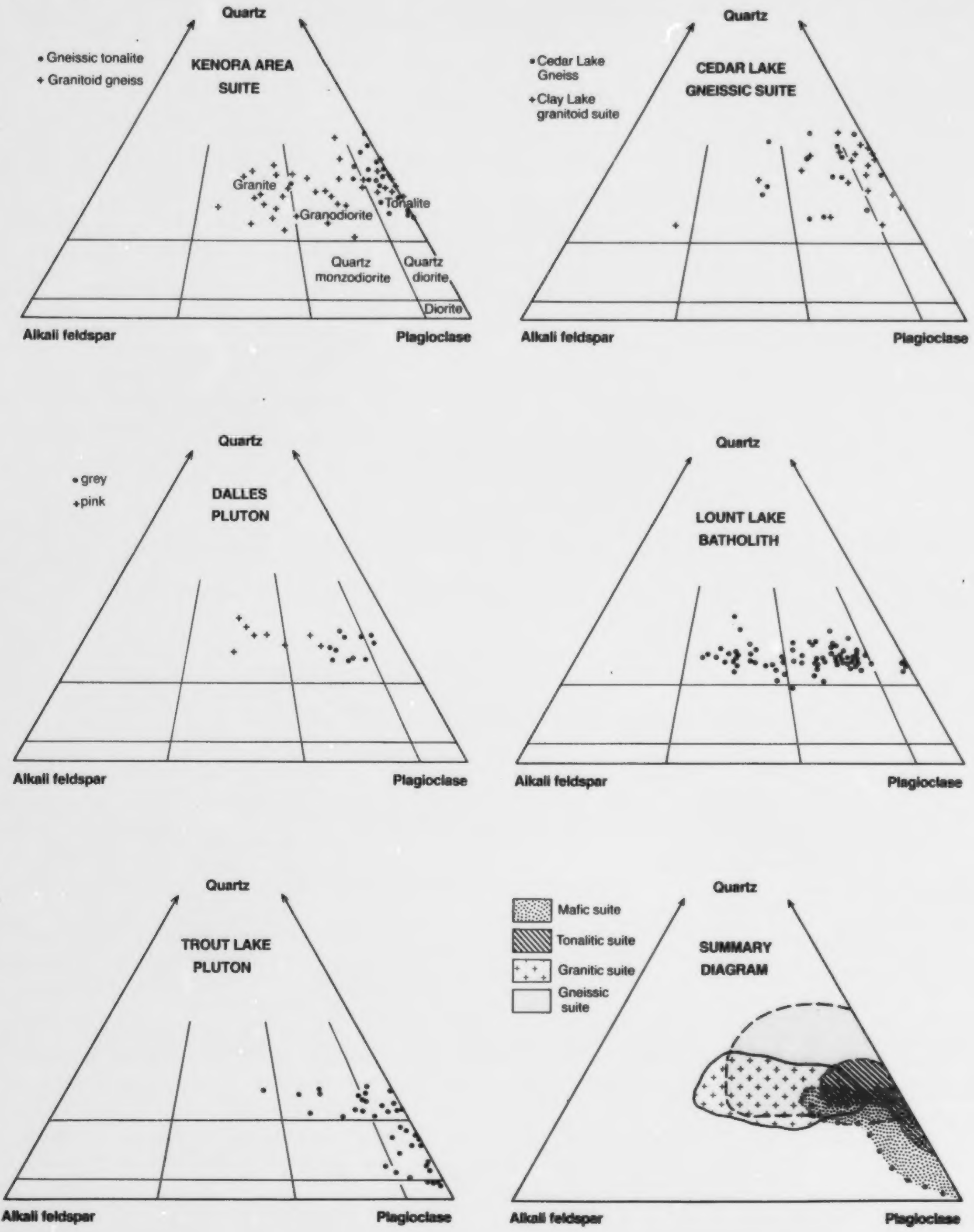


Figure 8.5. Ternary Q-A-P diagrams illustrating the modal mineralogy of Winnipeg River Subprovince granitoid rocks (data *from* Westerman 1977; Gower 1978; Beakhouse 1983).

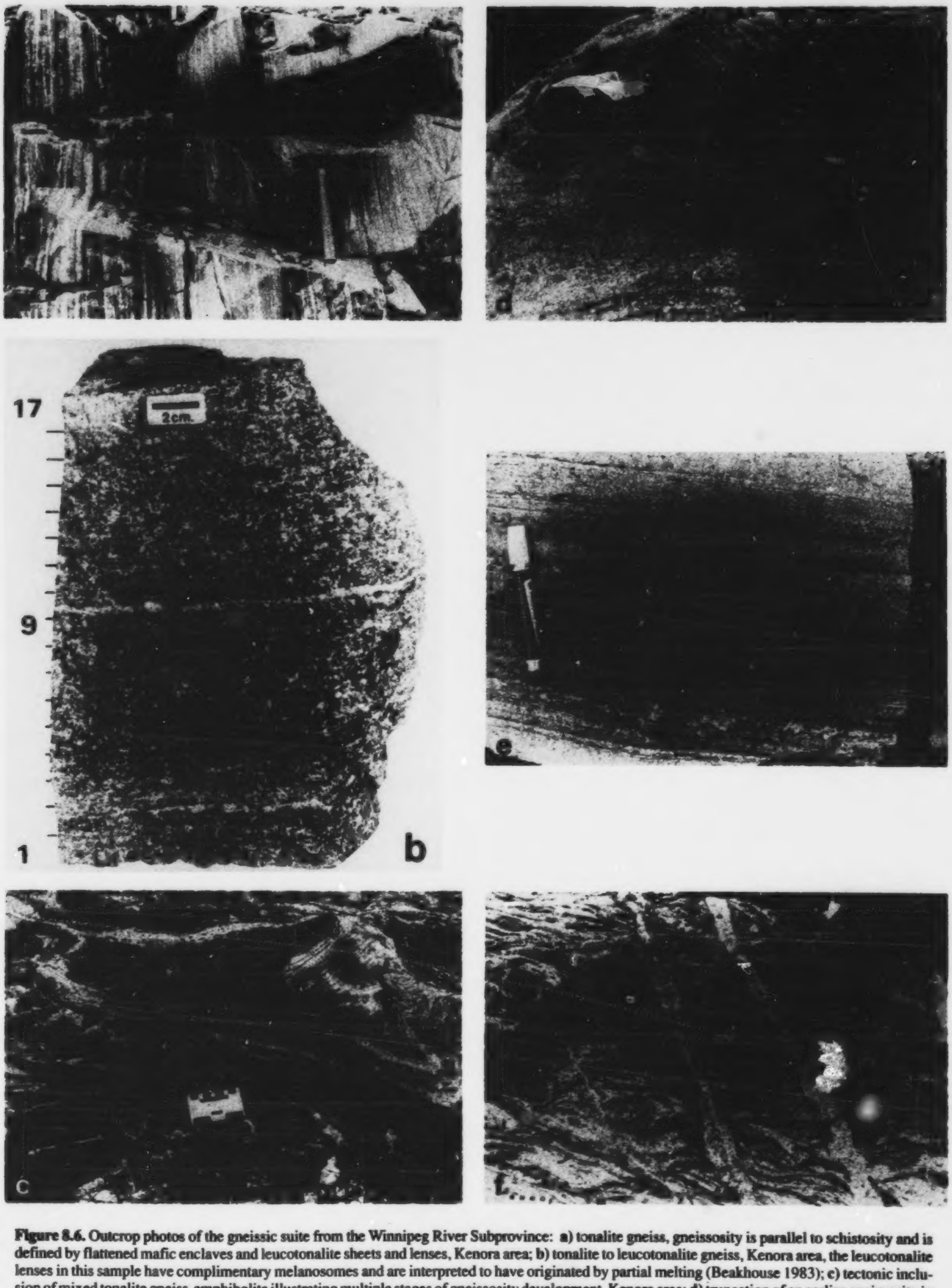


Figure 8.6. Outcrop photos of the gneissic suite from the Winnipeg River Subprovince: **a)** tonalite gneiss, gneissosity is parallel to schistosity and is defined by flattened mafic enclaves and leucotonalite sheets and lenses, Kenora area; **b)** tonalite to leucotonalite gneiss, Kenora area, the leucotonalite lenses in this sample have complimentary melanosomes and are interpreted to have originated by partial melting (Beakhouse 1983); **c)** tectonic inclusion of mixed tonalite gneiss-amphibolite illustrating multiple stages of gneissosity development, Kenora area; **d)** truncation of an earlier gneissosity in a rotated block of gneissic tonalite by a second gneissosity in granitic gneiss, Kenora area; **e)** microcline megacrysts and seams in gneissic tonalite near the margin of the Dalles pluton, Kenora area; and **f)** an example of a mixed rock (gneissic tonalite + granitic gneiss + amphibolite) cut by late zoned granitic pegmatite dikes, Cedar Lake area.

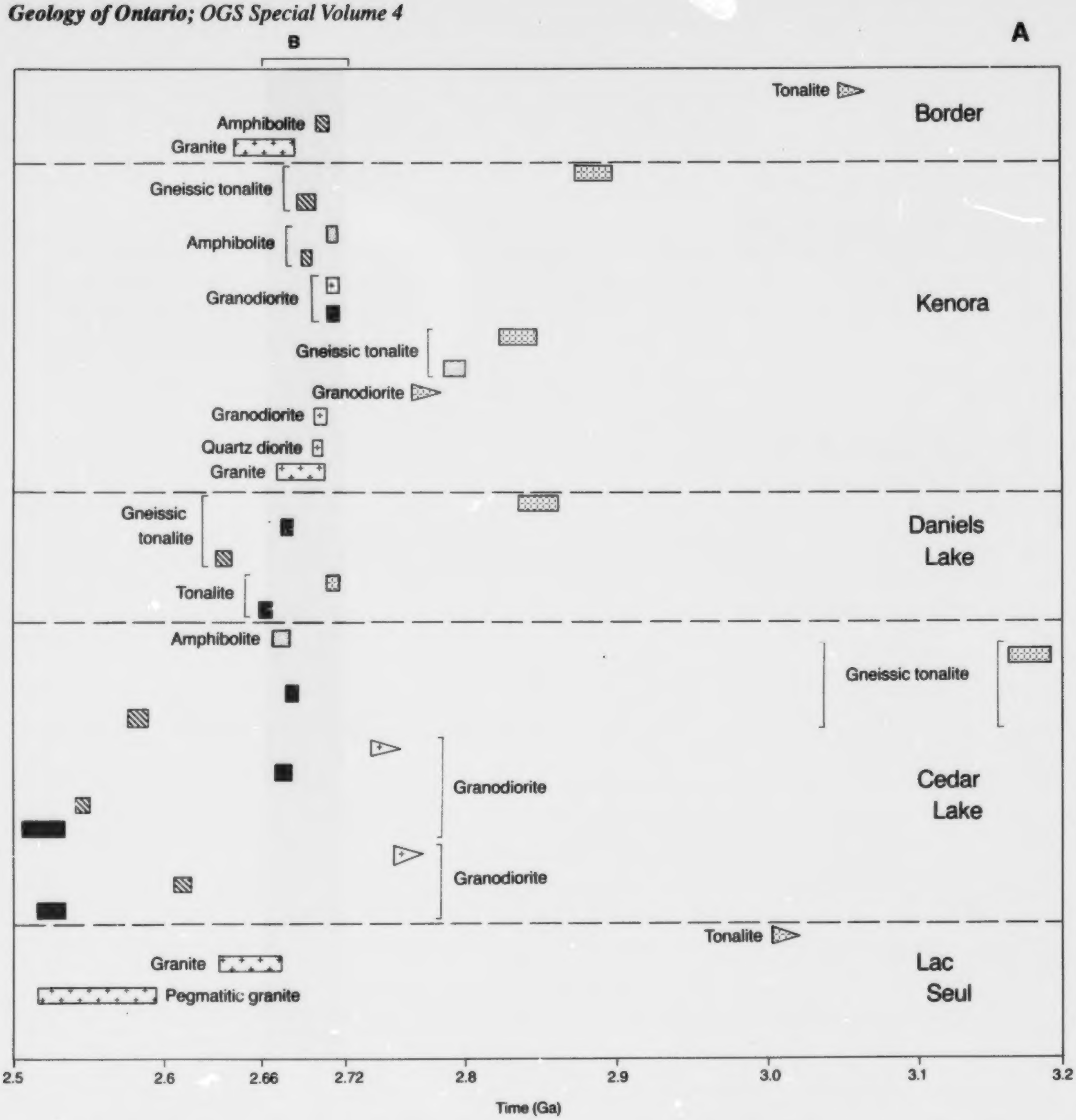


Figure 8.7. Summary of U-Pb geochronologic data for the Winnipeg River Subprovince. Part B is an enlargement of the area between 2.72 Ga and 2.66 Ga on Part A (data *from* Krogh et al. 1976a, 1976b; Beakhouse 1983; Beakhouse et al. 1988; Corfu 1988; Davis et al. 1988).

3. Some of the pegmatitic rocks consist largely of perthitic microcline with little or no quartz (Figure 8.6e) and are not likely magmatic. These rocks may originate from metasomatic fluids (e.g., microcline megacrysts adjacent to the Dalles Batholith, Beakhouse 1983).

Although the rocks of the granitoid gneiss component have diverse origins, the widespread occurrence of features indicative of *in situ* partial melting suggests that most were formed in, or emplaced into, rocks undergoing high-grade metamorphism. The nature of the granitoid gneiss component attests to the introduction of a hydrous fluid phase, with or without associated magmatic phases, which promoted partial melting and/or metamorphic differentiation of a range of rock types.

The individual components of the gneissic suite (amphibolite, gneissic tonalite and granitoid gneiss) each display internal variability and their mixture at various scales and proportions has produced complex polygenetic gneissic migmatitic rock (Figure 8.6f). This complexity is further enhanced by variable, but commonly intense strain and the emplacement of posttectonic dikes related to the granitic plutonic suite. Nevertheless, in most cases, these rocks can be explained by emplacement of tonalitic intrusions into amphibolite and the subsequent emplacement and/or *in situ* derivation of a more granitic phase.

Geochronological and isotopic investigations of the gneissic suite have documented a wide range of ages (*see* Figure 8.7). As discussed above, the age of the supracrustal

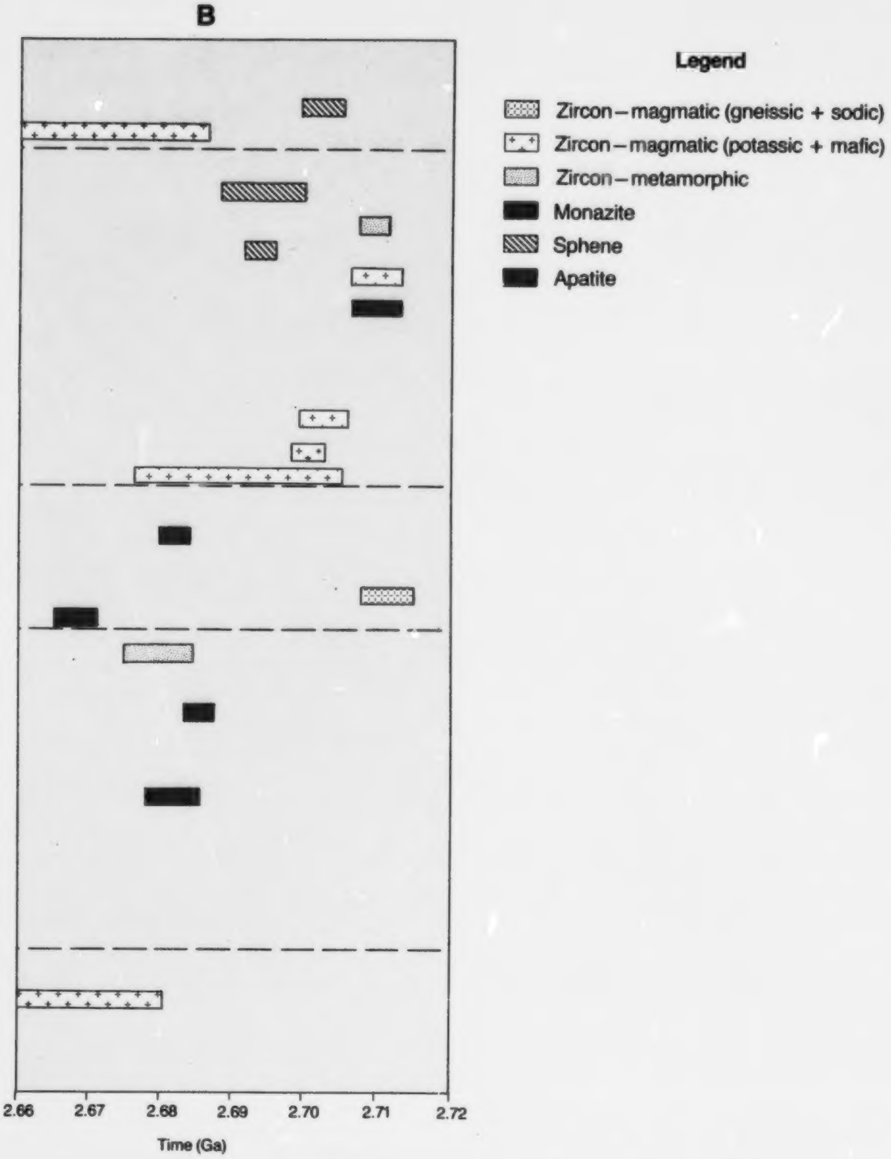


Figure 8.7. Summary of U-Pb geochronologic data for the Winnipeg River Subprovince. Part B is an enlargement of the area between 2.72 Ga and 2.66 Ga on Part A (data *from* Krogh et al. 1976a, 1976b; Beakhouse 1983; Beakhouse et al. 1988; Corfu 1988; Davis et al. 1988).

remnants within the gneisses has not been determined directly, and the age of the remnants is constrained only by the fact that they are intruded by gneissic tonalite. Most investigations have focussed on these foliated to gneissic tonalitic rocks and have documented a wide range of U-Pb ages (2835 to 3170 Ma; Krogh et al. 1976b; Beakhouse 1983; Corfu 1988; Davis et al. 1988). Several Rb-Sr investigations have suggested comparable, though less precise ages (Wooden and Goodwin 1980; Clark et al. 1981; Beakhouse 1983). The age of the granitoid gneiss component is not well constrained. This component is younger than the gneissic tonalite and older than some of the granitic and mafic plutons (2660 to 2705 Ma, *see* discussion below). The age of high-grade metamorphism for which a range of U-Pb ages are reported (2680 to 2789 Ma; Beakhouse 1983; Corfu 1988; Davis et al. 1988) provides a further general constraint on the age of the granitoid gneiss if the emplacement/generation of the granitoid gneiss accompanies high-grade metamorphism. The only example of granitic gneiss investigated isotopically are two leucotonalitic sheets from the Cedar Lake area (Beakhouse 1983). Limited U-Pb zircon data were inconclusive with the least discordant analyses having $^{207}Pb/^{206}Pb$ ages between 2.72 and 2.75 Ga. The highly radiogenic strontium of these whole rock samples suggests that, if they are as young as is suggested by the U-Pb data, they represent the products of melting of significantly older sialic crust (Beakhouse 1983).

Tonalitic Suite

The tonalitic plutonic suite is broadly similar to the gneissic tonalite component of the gneissic suite with respect to mineralogy, chemistry and possibly age. The tonalitic suite is distinguished from the gneissic suite by its generally more uniform, less complex, nonmigmatitic character (Breaks et al. 1978). These rocks characteristically have a foliation and/or lineation defined by the orientation of minerals, mineral aggregates and rare enclaves (Figure 8.8a). A weak gneissic fabric defined by narrow lenses of leucotonalite and discontinuous mafic schlieren is present locally, especially near the margins of plutons (Figure 8.8b).

The rocks of this suite are predominantly tonalite, granodiorite and, less commonly, quartz diorite (*see* Figure 8.5). Although some of these rocks plot in the granodiorite field they tend to be potassium-feldspar poor (modal abundance generally less than 15%). More potassium-feldspar-rich granodiorite to granite within some of these plutons may be products of the fractional crystallization of the tonalitic rocks (e.g., the pink phase of the Dalles pluton, Gower 1978), or may be younger anatectic derivatives of the more sodic rocks; but these phases are volumetrically minor and are not considered further in the following discussion.

These rocks are commonly leucocratic with a colour index (biotite and/or hornblende) typically less than 12%. Clinopyroxene is rarely preserved in the cores of some mafic clots. Microcline typically occurs as interstitial anhedra and patchy replacement of plagioclase. Minor amounts of epidote, magnetite, chlorite, sphene, zircon and apatite are typically present.

Enclaves are only abundant near the margins of some plutons. The great majority of enclaves are mafic to ultramafic blocks and schlieren composed principally of hornblende, plagioclase and biotite (after hornblende). Those enclaves that occur abundantly at the margins of plutons generally have blocky or tabular shapes and are similar to amphibolites occurring in the gneissic suite. These likely reflect incorporation of country rock near the level of emplacement. The less abundant, but more widely distributed, lenticular or wispy enclaves occurring throughout the plutons tend to have a higher colour index, and are interpreted to have been incorporated at greater depth. Some tabular mafic units in the Dalles pluton may be disrupted parts of a minor suite of calcalkaline dikes in the Kenora area (Gower 1978).

Relationship Between and Origin of the Tonalitic Suite and Tonalitic Component of the Gneissic Suite

Several lines of evidence suggest a close genetic and temporal relationship, notwithstanding the differing characteristics of the tonalitic plutonic suite and the gneissic tonalite component of the gneissic suite. The foremost evidence in favour of a close genetic and temporal relationship between these 2 groups of rocks is the fact that they commonly grade imperceptibly into one another (Breaks et al. 1978). Although the 2 form mappable units, in many cases, the tonalitic pluton becomes more strongly gneissic and contains a greater percentage of enclaves as the contact is approached. In such cases, the gneissic tonalite is often more melanocratic than the tonalitic pluton, but there is generally a continuum of mineralogical compositions.

Comparatively few geochronological investigations of the tonalitic suite have been carried out. The Sen Bay Complex, with a U-Pb zircon age of 3040±40 Ma (Krogh et al. 1976b), was described as a tonalite gneiss (Harris and Goodwin 1976), but is regarded as a tonalitic pluton by Breaks et al. (1978). The Dalles pluton has a comparatively young Rb-Sr whole rock age (2630±10 Ma, Gower and Clifford 1981), but a minimum ($^{207}Pb/^{206}Pb$) U-Pb zircon age of 2762 Ma for analyses that are sufficiently discordant to permit an age of crystallization comparable to that of the enclosing gneisses (Beakhouse 1983). Although more geochronological data are clearly required, the existing data are consistent with the interpretation that the tonalitic plutonic suite is significantly older than the other plutonic suites and is broadly comparable in age to the gneissic tonalite component of the gneissic suite.

The tonalitic components of the gneissic suite and the tonalitic suite may have a common origin because they have similar chemical characteristics, gradational field and mineralogical characteristics and comparable ages. The similar chemical characteristics include: high Na/K, CaO, Al_2O_3, low rubidium and initial strontium isotopic compositions, fractionated REE patterns with negligible europium anomalies and near-chondritic HREE (Gower et al. 1982, 1983). As discussed by Gower (1978), Gower et al. (1982, 1983) and Beakhouse and McNutt (1986, 1991), these characteristics are compatible with derivation of these rocks from the partial melting of metamorphosed tholeiitic basalt (mafic garnet granulite or quartz eclogite) at lower crustal or upper mantle depths.

Granitic Suite

The granitic suite underlies 39% of the Winnipeg River Subprovince (Breaks et al. 1978). Much of this suite occurs in several large batholiths (e.g., Lount Lake and Tetu batholiths) in addition to widely distributed smaller plutons. Numerous, widespread dikes of equigranular granite and granitic pegmatite and aplite are probably related to the granitic suite.

Field relationships indicate that the emplacement of the granitic suite is a late event in the history of the Winnipeg River Subprovince. The granitic suite intrudes the gneissic and tonalitic suites. The preservation of primary mineralogy, scarcity of textures indicative of recrystallization and lack of superimposed tectonic fabrics suggest that emplacement either postdates or is synchronous with the latest stages of the youngest episode of regional metamorphism and deformation. The mafic suite is similarly younger than these other events, but its relationship to the granitic suite is unclear. This interpretation is consistent with U-Pb geochronology, which is discussed more fully below.

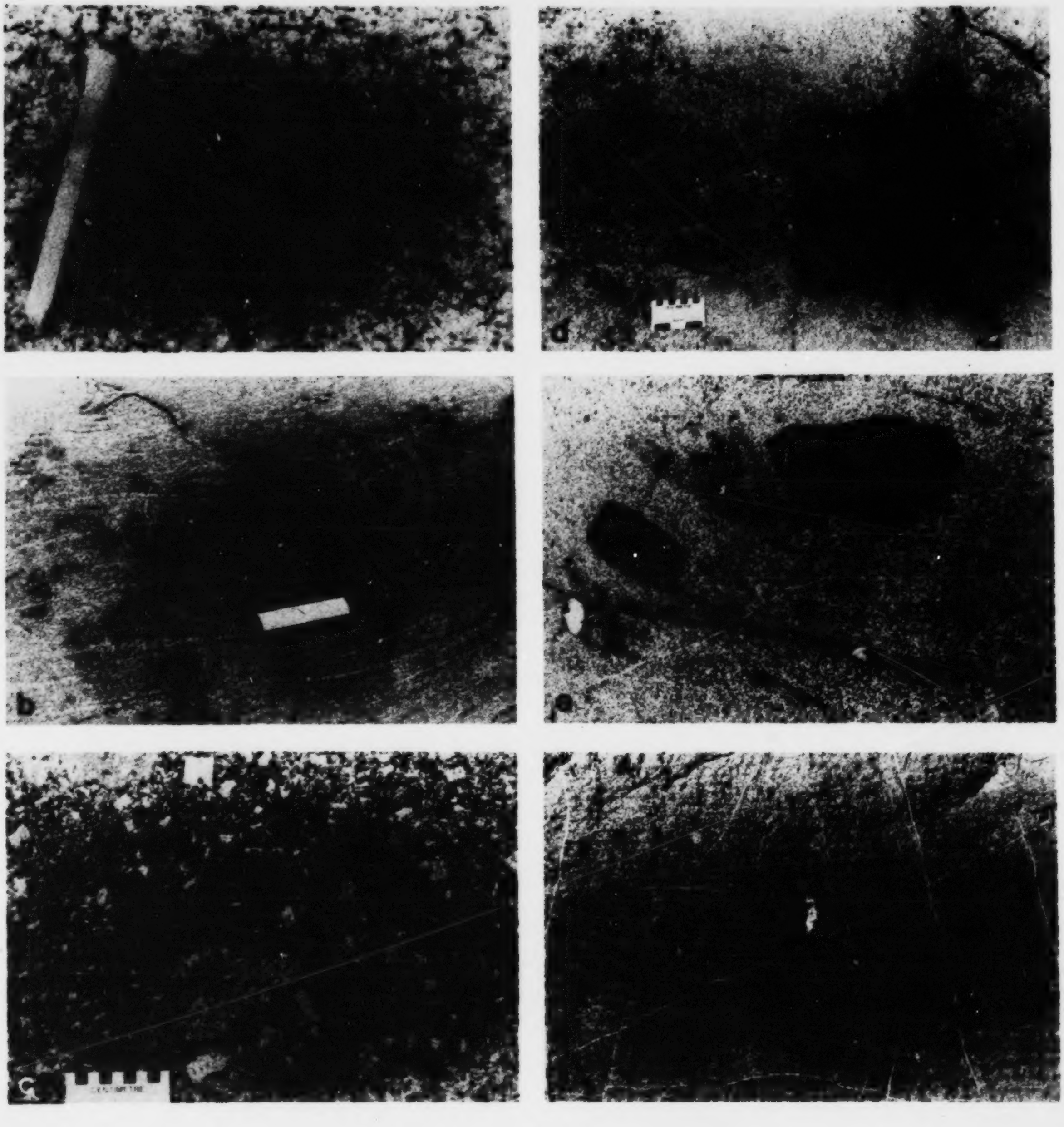


Figure 8.8. Outcrop photos of tonalitic, granitic and mafic suite plutonic rocks: **a)** moderately foliated tonalite, Dalles pluton; **b)** strongly foliated, weakly gneissose fabric near the contact of the Dalles pluton; **c)** granodiorite with well-developed, microperthitic, microcline megacrysts, Redditt area; **d)** medium-grained, equigranular granite with diffusely bounded, irregularly shaped, coarser grained pegmatitic "patches", Birch Lake area; **e)** amphibolitic and ultramafic enclaves in quartz diorite, Pelicanpouch pluton; and **f)** thinly layered hornblendite and gabbro (possible cumulate rock), Pelicanpouch pluton.

The granitic suite is represented by subequal abundances of granodiorite and granite together with volumetrically minor tonalite (Breaks et al. 1978). The tonalitic rocks are either inclusions of the tonalitic suite, or are plagioclase-rich cumulate phases related to the crystallization of the granodiorites (Beakhouse 1983). The most common enclaves within the granitic suite are sharply to diffusely bounded, commonly ghost-like, tonalitic gneisses with subordinate amphibolite. Enclaves are abundant in some granitic phases and are rare in the granodioritic phases.

The various phases of the Lount Lake batholith are typical of the granitic suite. The most abundant rock type is medium-grained granodiorite commonly characterized by large (less than 10 cm in diameter; usually approximately 1 to 3 cm in their longest dimension), subhedral to euhedral, perthitic microcline megacrysts (Figure 8.8c). On the basis of their textures, composition and phase equilibria considerations, these megacrysts have been interpreted to be late magmatic megacrysts rather than true phenocrysts or porphyroblasts (Beakhouse 1983). Biotite is the ubiquitous and dominant mafic phase. Subordinate hornblende occurs very locally in close proximity to amphibolitic enclaves and is interpreted to be xenocrystic. Common primary accessory phases include magnetite, epidote, sphene, zircon and apatite. Secondary alteration products include white mica, epidote and carbonate (after plagioclase), and chlorite and magnetite (after biotite).

The granites differ from the granodiorites in that biotite, in addition to ubiquitous, coarse plates of white mica that are distinct from the finer grained, shreddy deuteric alteration product of feldspar may be a primary mineral phase. Hornblende is never present. The secondary mineralogy is similar to the granodiorites except that epidote and carbonate are proportionately less abundant, reflecting the more albitic character of the plagioclase. The texture of the granites exhibits more variety than the granodiorites with both medium- and coarse-grained equigranular, pegmatitic and aplitic textures (Figure 8.8d) being common.

The rocks of the granitic suite are geochemically distinct from other plutonic rocks in the Winnipeg River Subprovince with high K_2O, rubidium, LREE abundances, variable, but generally high, HREE abundances, and prominent europium anomalies. Initial isotopic compositions of strontium (relatively radiogenic) and neodymium (relatively unradiogenic) indicate a crustal source for these rocks. These observations, together with field, petrologic, geochronological and other geochemical constraints, have led to the interpretation that the granitic suite is produced by the partial melting of the gneissic and tonalitic suite at mid- and lower-crustal depths (Beakhouse 1983; Beakhouse and McNutt 1986, 1991).

Mafic Suite

The mafic suite underlies less than 2% of the Winnipeg River Subprovince (*see* Figure 8.2). This description is based primarily on the pluton which is referred to variously as the Pelicanpouch pluton (Breaks et al. 1978), the Trout Lake pluton (Beakhouse 1983) and the Muriel diorite-granodiorite (Gower 1978). Field investigations of other examples (e.g., Ena pluton) suggest that similar rock types and relationships characterize this suite.

The Pelicanpouch pluton is an elongate, composite intrusion that intrudes the gneissic suite. Based on its undeformed, unmetamorphosed character, the pluton is interpreted to be one of the younger suites. This is consistent with a U-Pb zircon age for a quartz diorite from the Pelicanpouch pluton (2700±2 Ma; Beakhouse et al. 1988) which slightly postdates regional metamorphism in the area, and is within the range of ages reported for the granitic suite.

Rocks of the Pelicanpouch pluton range in composition from ultramafic to granitic but most are quartz diorite, tonalite and granodiorite (*see* Figure 8.5). Granitic rocks are rare dike phases that may be unrelated to the pluton and are not considered further. Amphibolite enclaves are widespread in all phases of the pluton (Figure 8.8e). Ultramafic rocks occur as small (less than 20 cm), widely distributed, hornblende ± biotite clots and, less commonly, as larger blocks of gabbro-hornblendite within which layering is preserved (Figure 8.8f). On the basis of this textural evidence, these rocks are interpreted to be ultramafic cumulates complementary to the enclosing more felsic rocks (Gower 1978; Beakhouse 1983). These ultramafic enclaves occur within all phases, but are most abundant in the dioritic to quartz dioritic phases.

Quartz diorite, tonalite and granodiorite, together with minor quartz monzodiorite, diorite and gabbro, form a compositional and textural continuum. In addition to the varying proportions of quartz and the feldspars, there is a progressive change in colour index from dioritic and quartz dioritic phases (colour index 20 to 40), tonalitic phases (colour index 10 to 20) to granodioritic phases (colour index approximately 10) (Beakhouse 1983). In spite of the variation in the proportions of the major rock-forming minerals, certain petrographic aspects are common to all phases: 1) hornblende predominates over biotite; 2) presence of magmatic epidote; 3) accessory minerals include abundant sphene along with magnetite, zircon and apatite; and 4) similar secondary mineralogy (epidote, white mica, carbonate, chlorite) in proportions determined by those of the primary mineralogy. Petrographic dissimilarities include: 1) the occurrence of microcline as small interstitial grains in the more plagioclase-rich phases in contrast to granodioritic phases where it occurs mainly as subhedral, late magmatic megacrysts up to 3 cm in diameter; and 2) the restriction of rare allanite to the granodioritic phases.

Where intrusive relationships are unambiguous they suggest that there is a general relative age progression with more melanocratic, plagioclase-rich phases followed by progressively more leucocratic and potassium-enriched phases. However, adjacent phases commonly exhibit gradational contacts or, rarely, have mutually intrusive relationships most commonly manifested as spheroidal enclaves of 2 adjacent phases within each other near their contact. Emplacement of the various phases is considered to have been broadly synchronous.

The lithological diversity of the pluton is reflected in the geochemistry of the main phase groups with SiO_2

ranging from a low of 53.3% in the diorites to a high of 71.6% in the granodiorites (Beakhouse 1983). The tonalitic and granodioritic phases exhibit smooth, continuous variation of major and several trace elements, suggesting that they are genetically related. The dioritic phases plot as somewhat more diffuse continuations of these trends for many elements; but there are noteworthy discontinuities in Al_2O_3, zirconium, strontium (lower in diorites), K_2O and rubidium (higher in diorites). These discontinuities suggest that 2 distinct magma sources are present and that these cannot be related by crystal fractionation (Beakhouse 1983).

The initial Sr isotopic composition of the dioritic phases is 0.7021±.0003 (Beakhouse 1983). Wooden and Gower (*cited in* Gower and Clifford 1981) reported a slightly different value of 0.7016±.0001 (based on analyses of dioritic to granodioritic phases, but is largely constrained by the diorite data). The significance of these initial ratios is difficult to evaluate because of the uncertainty in the nature of the source and its Rb:Sr ratio.

STRUCTURAL GEOLOGY

The Winnipeg River Subprovince is characterized by strongly deformed, heterogeneous, gneissic units (supracrustal and gneissic suites) that wrap around less deformed, more homogeneous units (tonalitic, granitic and mafic suite). The former occur most abundantly as moderately to steeply dipping units along the southern boundary of the subprovince. The strike of gneissosity, within a zone (approximately 200 to 2000 m wide) adjacent to the subprovincial boundary, is parallel to the boundary. At a short distance north of the boundary, the dips typically become moderate to shallow and the units have a more arcuate geometry reflecting the second order control of more massive plutons, which are cores to folds and structural domes.

Deformed rocks in the Winnipeg River Subprovince display a wide range of mesoscopic structural complexity due to: the relative timing of emplacement and deformation; relative ductility and geometric form of the units; the presence of suitable markers that can record the deformation; and regional variation in the intensity of strain. In general, the most complex structure(s) and highest degree of strain are displayed by the gneissic suite and compositionally heterogeneous supracrustal remnants. The same deformational history was experienced by the more homogeneous supracrustal remnants such as massive, fine- to medium-grained basalt, but their lack of strain markers and simple mineralogy makes this less apparent. Similarly, evidence of a complex deformational history within the tonalitic suite is noted locally but, in many cases, and particularly in the larger plutons, the lack of strain markers and their differing response to the same stress that complexly deformed the more heterogeneous rocks makes this less apparent. Only rarely do the granitic and mafic suites display evidence of a superimposed tectonic deformation and their emplacement is interpreted to be either late tectonic or posttectonic.

Understanding of deformational processes in the Winnipeg River Subprovince is based primarily on the detailed analysis of specific areas (Westerman 1977; Gower 1978; Gower and Clifford 1981) together with more general reconnaissance investigation (Breaks et al. 1978; Beakhouse et al. 1983; Sanborn–Barrie 1988). Although there are numerous variations and exceptions, the data are broadly consistent with the sequential structural history summarized below.

The earliest deformation is recorded in the supracrustal remnants and some of the metaplutonic rocks. Near Kenora, tonalitic gneisses locally truncate leucogranitoid veins and a foliation in the supracrustal remnants (Gower and Clifford 1981), suggesting that the earliest deformation precedes at least some tonalitic plutonism. The earliest widespread deformational structures are mesoscopic, isoclinal to tight, commonly rootless, intrafolial folds with axial surfaces parallel to the plane of the gneissosity (Westerman 1977; Gower and Clifford 1981; Gower 1978). Locally, there is evidence that these are refolded by tight to isoclinal folds with similarly oriented axial surfaces (Westerman 1977). The intense fabric development, flattening, elongation and transposition of a lithologically heterogeneous assemblage associated with this early deformation is largely responsible for the strongly banded character of these rocks. The timing of this deformation is (for the most part) post-tonalite emplacement and pre-doming (*see* discussion below). The origin of these structures has been tentatively correlated with large-scale horizontal tectonic structures (thrusts and/or nappes; Westerman 1977) that have similarly constrained relative ages.

At Cedar Lake, in the core of a number of structural domes, paragneiss structurally underlies metaplutonic rocks (Figure 8.9a). The latter was interpreted, on the basis of comparison of structural histories as well as regional considerations, to be older than the paragneiss which presently underlies it (Westerman 1977). Although the age of the paragneiss remains uncertain, subsequent U-Pb geochronological investigations have demonstrated that the overlying tonalitic gneiss (3170 Ma, Corfu 1988) is the oldest rock heretofore dated in the western Superior Province. The age relationships require that either the entire sequence at Cedar Lake represents the lower limb of a large-scale recumbent fold or that some or all of the units are allochthonous thrust slices (Westerman 1977). Further evidence for large-scale, subhorizontal tectonics is afforded by a large-scale, recumbent fold structure near Silver Lake (Figure 8.9b; Beakhouse et al. 1983). The geometry of this fold structure is closely comparable to that of the mesoscopic folds discussed above and is consistent with the suggestion that some of these minor folds are related to large-scale tectonic transport (Westerman 1977). The timing of development of the recumbent folding and subsequent doming in the Silver Lake area is tightly constrained by the age of the deformed Daniels tonalite (2710^{+5}_{-2} Ma, Corfu 1988) and the postdoming Lount Lake batholith (2702^{+3}_{-4} Ma, Beakhouse 1983; Beakhouse et al. 1988).

The next stage in the structural development of the Winnipeg River Subprovince is the deformation of the tabular, strongly deformed units together with their contained minor structures and bounding faults into upright, open,

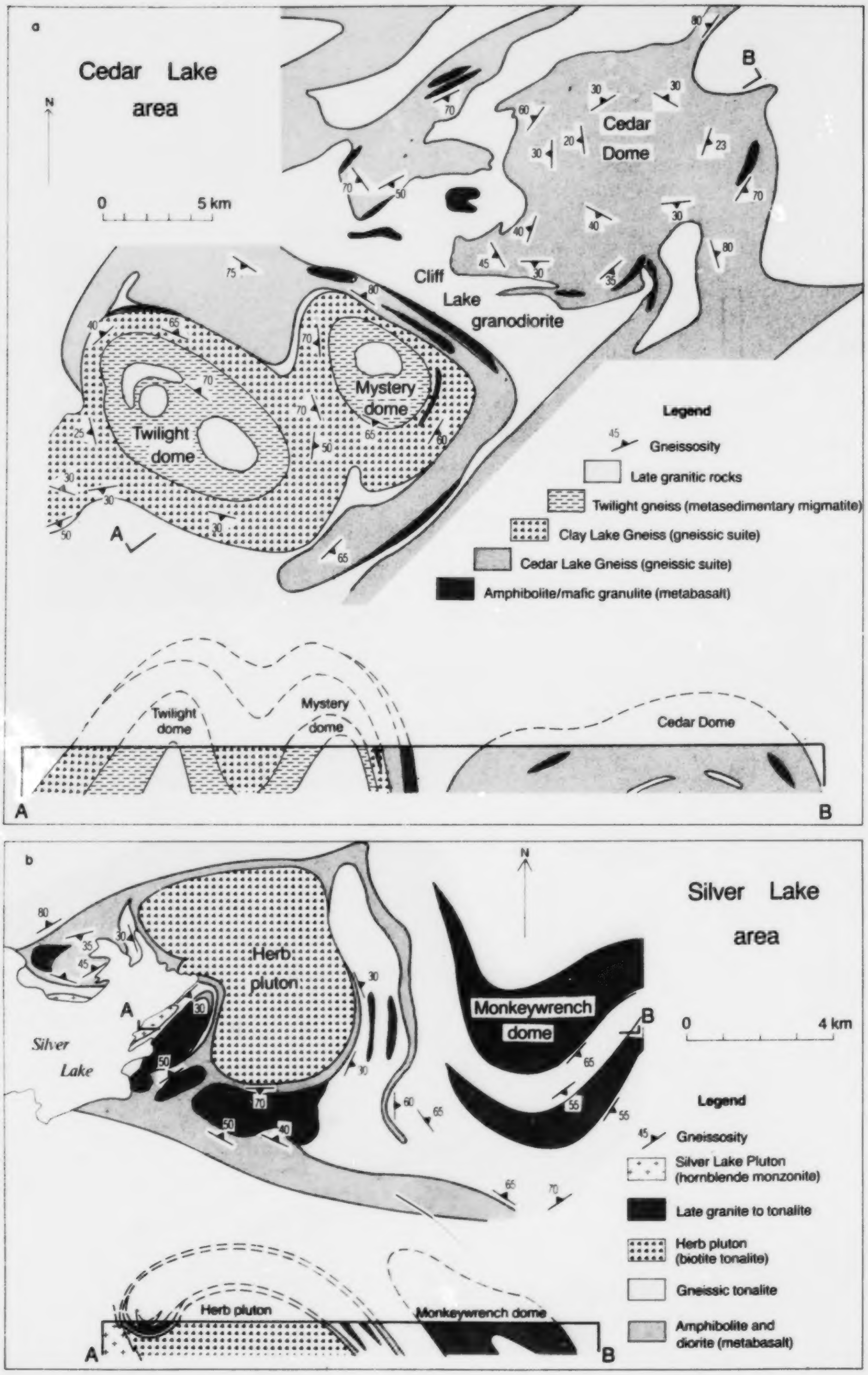

a
Cedar Lake area
N
0
5 km
Cedar Dome
Cliff Lake granodiorite
Twilight dome
Mystery dome
A
B
Legend
Gneissosity
Late granitic rocks
Twilight gneiss (metasedimentary migmatite)
Clay Lake Gneiss (gneissic suite)
Cedar Lake Gneiss (gneissic suite)
Amphibolite/mafic granulite (metabasalt)
Twilight dome
Mystery dome
Cedar Dome
A
B
b
Silver Lake area
N
Herb pluton
Monkeywrench dome
Silver Lake
A
B
0
4 km
Legend
Gneissosity
Silver Lake Pluton (hornblende monzonite)
Late granite to tonalite
Herb pluton (biotite tonalite)
Gneissic tonalite
Amphibolite and diorite (metabasalt)
Herb pluton
Monkeywrench dome
A
B

Figure 8.9. Geological maps with cross-sections of **a)** the Cedar Lake area (*modified from* Westerman 1977) and **b)** the Silver Lake area (Beakhouse et al. 1983).

commonly domical folds. Some of these (e.g., Dalles Batholith, Gower 1978; Gower and Clifford 1981; Herb pluton, Beakhouse et al. 1983) are mantled gneiss domes. The emplacement of the Dalles pluton is responsible for enhancement of gneissosity adjacent to the batholith, development of a rim synform and displacement along faults (Gower 1978; Gower and Clifford 1981). Final upward emplacement that produced these structures may have been as a solid diapir that postdates the crystallization of the batholith. Even those domes without an exposed core probably reflect diapiric emplacement of underlying granitoid rocks (Westerman 1977; Shanks 1985).

The final major deformation affecting the Winnipeg River Subprovince is transcurrent shearing and faulting mainly at or near the subprovince boundaries. In some cases, this deformation represents reactivation of faults that were initiated earlier (Gower 1978). Most commonly, shear zones have an easterly strike and a dextral sense of displacement or have a more northeasterly strike (0 to 60° azimuth) and a sinistral sense of displacement. This is consistent with a regional transcurrent fault pattern first noted by Schwerdtner et al. (1979) reflecting late northwesterly directed regional shortening.

Faults at the subprovince boundary in the vicinity of Kenora and Sioux Lookout record a final stage of north-side-up, dip-slip movement on near-vertical faults. This movement is interpreted to reflect isostatic compensation for tectonic thickening of the crust to the north of the boundary.

Deformation within the subprovince bounding faults persisted to the latest Archean based on the observation that all Archean rocks are deformed in these zones and that the 2.12 Ga Kenora–Kabetogama dikes (Southwick and Day 1983) are undeformed in these zones. Gower (1978) reported that some minor faulting postdates the Proterozoic dikes.

In summary, the structural evolution of the Winnipeg River Subprovince is interpreted to consist of an early phase of horizontal tectonics manifested as thrust faults and recumbent, isoclinal fold structures that were subsequently deformed in a later tectonic phase of upright folding, doming and predominantly dip-slip movement on near-vertical faults. The timing and cause of these structures are discussed in a subsequent section. A phase of transcurrent faulting concentrated near the margin of the subprovince is a manifestation of latest Archean northwesterly directed regional compression that affected the entire western Superior Province. Subsequent dip-slip movement on some of these faults may be related to isostatic adjustments related to the final stabilization of the Superior Province.

METAMORPHISM

The Winnipeg River Subprovince is characterized by medium- to high-grade metamorphism affecting the supracrustal remnants and belts, gneissic suite and at least some of the tonalitic plutonic suite although the evidence for the latter is more cryptic. The granitic and mafic plutonic suites are either synchronous with or postdate regional metamorphism. Although the evidence for this metamorphism is widespread, the precise peak metamorphic conditions and possible regional variation in grade are poorly understood due to the paucity of mineral assemblages that can tightly constrain P-T conditions.

Ubiquitous evidence for relatively high-grades of metamorphism includes the presence of: 1) *in situ* partial melting of a variety of rocks; and 2) mafic enclaves composed of plagioclase + hornblende + quartz ± clinopyroxene ± orthopyroxene. *In situ* partial melting giving rise to leucotonalite within amphibolite and leucotonalite to leucogranite in metatonalites is indicated by physically isolated lenses and pods of the more leucocratic phase coexisting with a complementary melanocratic rock (restite) within a host of appropriate bulk composition (Westerman 1977; Gower 1978; Beakhouse 1983). These observations suggest that metamorphic temperatures exceeded those required to melt tonalite and gabbro with excess water approximately 630° to 700°C at pressures in excess of 200 MPa (Wyllie 1977). The ubiquitous presence of hornblende and absence of chlorite and actinolite in mafic supracrustal enclaves indicates metamorphic temperatures in excess of 550° to 600°C (Liou et al. 1974). Clinopyroxene-hornblende amphibolites and orthopyroxene-clinopyroxene-hornblende mafic granulites characterize parts of the subprovince (Beakhouse et al. 1983) suggesting that temperatures in excess of 800°C were attained locally (Spear 1981).

More detailed investigations of metamorphic conditions are consistent with the generalizations outlined above. At Kenora, the mineralogy of gneissic tonalites and a variety of entrained supracrustal rocks have been interpreted to constrain peak metamorphic conditions to 650 to 750°C at 400 to 700 MPa (Gower 1978). In the eastern Lac Seul region, estimates of peak metamorphic conditions include 650 to 750°C at 350 to 600 MPa (Harris 1976), greater than 770°C at 400 to 600 MPa (Breaks et al. 1978; Breaks 1988) and 700 to 750°C at 400 to 600 MPa (Perkins and Chipera 1985; Chipera and Perkins 1988). Although many of the critical observations in the eastern Lac Seul area pertain to the metasedimentary gneisses located immediately north of the subprovince boundary, granulite-facies metamorphism straddles the subprovince boundary (Thurston and Breaks 1978; Breaks 1988). Granulite-grade metamorphism in the Cedar Lake area developed under conditions comparable to those interpreted for the eastern Lac Seul area (Westerman 1977).

The duration and number of metamorphic episodes within the Winnipeg River Subprovince is uncertain. Westerman (1977) postulated the existence of multiple episodes of high-grade metamorphism in the Cedar Lake area. The U-Pb ages of metamorphic zircons from mafic supracrustal rocks in the Kenora and Cedar Lake areas (2709±2 Ma and 2678^{+6}_{-4} Ma respectively; Corfu 1988) and zircons from a pegmatitic leucosome in paragneiss from the eastern Lac Seul area (2681±20 Ma; Krogh et al. 1976b) indicate that a comparatively late period of high-grade metamorphism affected the entire subprovince. It is unclear whether the younger metamorphic ages determined for the granulite-facies areas, as compared to that from the Kenora area, reflect discrete metamorphic events or slower cooling at the greater depths suggested by their mineralogy. The observation that U-Pb sphene ages are significantly younger in the Cedar Lake area than in the Kenora area (Corfu 1988) is consistent with the latter hypothesis. In the Kenora area, a gneissic tonalite with morphologically distinct zircons (2790^{+9}_{-7} Ma) interpreted to be related to incipient partial melting, is approximately 40 Ma younger than magmatic zircons from the same sample (Beakhouse 1983). This may record an example of the earlier metamorphic episodes postulated by Westerman (1977).

Evidence suggests that the entire Winnipeg River Subprovince underwent medium- to high-grade regional metamorphism subsequent to 2710 Ma, thereby postdating tonalitic plutonism and slightly predating and partially overlapping late tectonic to posttectonic granitic and mafic plutonism. There is less conclusive evidence for earlier episodes of high-grade metamorphism. Constraints on the conditions of metamorphism imply that the present erosional surface has been exhumed from a depth of 14 to 21 km and that average geothermal gradients were approximately between 30 and 60°C/km. It is probable that the horizontal tectonic regime immediately preceding regional metamorphism resulted in the creation of transient geothermal gradients (England and Thompson 1986), and the estimates above are undoubtedly an oversimplification of the thermal structure of the crust.

GEOLOGICAL HISTORY

A synthesis of the observations and interpretations discussed above leads to a general geological history of the Winnipeg River Subprovince as follows. Prior to 2.83 Ga, crust evolved in a two-stage process involving production of large amounts of tholeiitic basalt from the mantle and the melting of much of this basalt to produce magmas of tonalitic affinity. This was followed by a period of no recorded activity that persisted for approximately 100 Ma, and coincided with an interval of widespread volcanism and plutonism in the adjacent western Wabigoon Subprovince. At approximately 2710 Ma, coincident with the cessation of widespread magmatic activity in the western Wabigoon Subprovince, the crust of the Winnipeg River Subprovince was tectonically thickened by large-scale thrusting and nappe emplacement. Thickening initiated both the development of domes in response to induced gravitational instability and high-grade metamorphism as a consequence of deep burial. A further consequence of the high-grade metamorphism was extensive intracrustal melting of dominantly tonalitic crust giving rise to a distinctive suite of granodioritic to granitic plutonic rocks emplaced over approximately 40 Ma. A protracted period of slow crustal cooling accompanied and outlasted this granitic plutonism. These relationships, together with the constraining U-Pb ages, are summarized in Figure 8.10. A more detailed discussion of critical evidence for the nature and origin of the different stages of this sequence of events and their interrelationships follows.

The earliest recognized stage in the evolution of the Winnipeg River Subprovince is the deposition of supracrustal rocks preserved as highly disrupted enclaves in tonalitic metaplutonic rocks. This sequence was deposited subaqueously and consists mainly of metamorphosed tholeiitic basalt with minor chemical metasedimentary rocks. The predominance of extensive basaltic volcanism is further suggested by the interpretation that many of the subsequently emplaced tonalites originate by the partial melting of basalt.

The supracrustal sequence is intruded by tonalitic rocks that are older than 2.83 Ga and may be further divisible into 2.83 to 2.9 Ga and 3.0 to 3.2 Ga groups that may represent discrete episodes of crust formation. Regardless of their age, these tonalites display close mineralogical and geochemical similarities to each other and tonalitic rocks from elsewhere. Noteworthy textural and mineralogical attributes include the predominance of fine- to medium-grained, equigranular textures, mafic phases consisting of biotite with or without hornblende and rare clinopyroxene and the absence of aluminous phases. Distinctive geochemical characteristics include high Na/K, calcium and strontium and low potassium, rubidium and initial strontium isotopic composition. REE patterns are strongly fractionated with near chondritic HREE and negligible europium anomalies. These characteristics are most readily explained by low degrees of partial melting of metamorphosed tholeiitic basalt (mafic garnet granulite or quartz eclogite).

Although the mafic volcanic component is volumetrically minor and poorly preserved, the crust of the Winnipeg

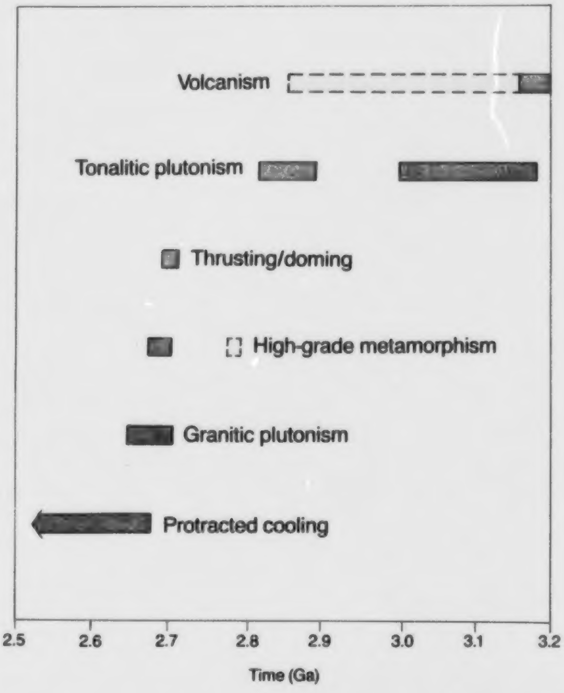


Figure 8.10. A generalized summary of the geological history of the Winnipeg River Subprovince. Relative and absolute chronology is based on the data presented in Figure 8.7 and relationships discussed in the text.

River Subprovince, prior to 2.83 Ga, is analogous to bimodal basalt-tonalite suites that are a recurring theme in Archean crustal evolution (Barker and Peterman 1974). This association has been interpreted to reflect a two-stage magma generation process whereby melting of the mantle produces tholeiitic basalt which is in turn partly melted, often within 50 Ma of its eruption, at lowermost crustal or uppermost mantle depths to produce magmas of tonalitic affinity (Hanson and Goldich 1972; Arth and Hanson 1972, 1975). Such an interpretation is broadly consistent with the early history of the Winnipeg River Subprovince.

The existence and nature of deformation or metamorphism, prior to 2.83 Ga, is speculative. The emplacement of large volumes of plutonic rock into the early supracrustal rocks implies significant deformation. A pre-tonalite foliation recognized in the Kenora area (Gower and Clifford 1981) is the only fabric unambiguously interpreted to have developed during this early period. Multiple episodes of metamorphism have been proposed and it is possible that some of these occurred prior to 2.83 Ga although there is no conclusive evidence that this is the case. A probable metamorphic zircon from a tonalite gneiss in the Kenora area has a U-Pb age of 2790^{+9}_{-7} Ma (Beakhouse 1983) suggesting that a period of high-grade metamorphism may punctuate the end of tonalite plutonism and the beginning of a quiescent interval.

There are few indications for geological activity within the interval between 2.71 and approximately 2.8 Ga. The only exceptions are from the Cedar Lake area where late tonalitic layers in the gneissic rocks have $^{206}Pb/^{207}Pb$ zircon ages between 2720 and 2750 Ma (Beakhouse 1983) and the Cliff Lake granodiorite has a complex array of $^{206}Pb/^{207}Pb$ zircon ages as old as 2756 Ma (Corfu 1988). In both cases the ages are poorly constrained and their interpretation is problematical. The probable interval of quiescence between 2.71 and 2.8 Ga coincides with the timing of widespread volcanism and plutonism in the adjacent western Wabigoon Subprovince (*see* Davis et al. 1988 and the references cited therein).

The next stage in the development of the Winnipeg River Subprovince involved formation of thrust and/or nappe-like structures, which were subsequently deformed into open domical structures. At Silver Lake, the recumbent nappe-like structure and subsequent doming is tightly constrained at 2707±8 Ma. The timing of development of these structures is not as well constrained elsewhere, but field relationships and geochronological data are broadly consistent with this interpretation. The timing of these nappe-like structures corresponds with the timing of thrust-stacking of supracrustal sequences in adjacent parts of the western Wabigoon Subprovince near Sioux Lookout (Davis et al. 1988). The implication is that at 2707±8 Ma, the Winnipeg River Subprovince and adjacent portions of the western Wabigoon Subprovince both underwent tectonic thickening. Furthermore, this is the earliest stage in their evolution that these 2 subprovinces appear to share a common history.

A third magmatic stage and a period of high-grade, regional metamorphism commenced immediately after or during tectonic thickening. The data suggest that peak metamorphic conditions were 630 to 800°C at pressures between 400 and 700 MPa. This in turn suggests that average geothermal gradients were between 30 and 60°C/km and that the present erosional surface has been exhumed from depths of 14 to 21 km. There is some regional variation in metamorphic grade from upper amphibolite to granulite-facies. The data suggest that areas of higher grade metamorphism, such as Cedar Lake, have slightly younger metamorphic zircons and significantly younger sphene ages than areas of somewhat lower grade metamorphism such as at Kenora. This may reflect slower cooling in deeper levels of the crust.

The third magmatic stage is represented by granodioritic to granitic batholiths and plutons emplaced between 2.66 and 2.71 Ga. These plutons are mineralogically distinct from the earlier tonalitic plutons and differ geochemically by virtue of higher abundances of potassium, rubidium and trivalent REE, lower calcium and strontium, variable but generally negative europium anomalies and radiogenic isotopic (strontium and neodymium) signatures indicative of crustal sources. These characteristics reflect their origin as partial melts of older, tonalitic crust. The timing of this plutonism with respect to tectonic thickening, their abundance, interpreted crustal origin and the absence of evidence for a mantle component suggests that the generation of these rocks is due primarily to deep burial.

Transcurrent deformation in narrow zones localized near the subprovince boundary locally deforms stage 3 plutons. It is probable that some transcurrent movement occurred on these and other faults earlier in their history as well. The latest movement on some of these faults involved a dip-slip component, which juxtaposed higher metamorphic grade rocks against those characterized by lower metamorphic grades.

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Chapter 9

Wabigoon Subprovince

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Abstract

The Wabigoon Subprovince is a 900 km long, east-trending granite-greenstone subprovince composed of metavolcanic and subordinate metasedimentary rocks which are surrounded and cut by granitoid batholiths. The subprovince is bounded on the north by the Winnipeg River and English River subprovinces and on the south by the Quetico Subprovince. The Wabigoon Subprovince has been divided into 3 regions differing in proportions of the major units and structural style: 1) the western region, characterized by broad expanses of supracrustal rocks with ovoid, synvolcanic, polyphase batholiths and minor gneissic units; 2) the central region, underlain by ovoid gneiss domes and elliptical batholiths with screens and small belts of metavolcanic rocks; and 3) the eastern region, composed of supracrustal rocks and lobate synvolcanic granitoid batholiths.

In the western region, supracrustal rocks are subdivided into lithostratigraphic units grouped into tectonically bounded assemblages. The assemblages can be classified into 4 major types of sequences : 1) lower mafic sequences composed of tholeiitic and minor komatiitic basalt with lesser amounts of andesite mafic flows, fragmental rocks and coeval intrusions; 2) diverse intermediate to felsic metavolcanic sequences composed of calc-alkalic rocks and lesser amounts of tholeiitic andesite to rhyolite pyroclastic rocks with subordinate epiclastic units and flows; 3) upper mafic sequences composed of komatiitic and tholeiitic mafic flows; and 4) metasedimentary sequences composed of turbiditic, alluvial-fan fluvial and minor platform sequences.

Granitic units comprise pre-greenstone granitoid plutons in the central region of the subprovince, numerous synvolcanic, polyphase batholiths and posttectonic stocks. Ultramafic intrusions are divided into those: coeval with mafic volcanism; coeval with felsic volcanism and early granitoid plutonism; coeval with late granitoid plutonism; and serpentinized, tectonically bounded units.

Geochronologic data indicate that the central part of Wabigoon Subprovince has a 3 billion-year-old granitoid basement and *circa* 3 billion-year-old greenstones. In the western Wabigoon, predeformation supracrustal sequences range from 2775 to 2718 million years old, and syndeformational metasedimentary sequences are less than 2714 to 2696 million years old.

The Wabigoon Subprovince has been subjected to at least 2 major structural events. These events are an early aggregation of supracrustal assemblages of similar or contrasting age followed by the interaction of the Wabigoon Subprovince with its neighbours, resulting in contrasting patterns between the interior and the margins of the subprovince. The interior consists of anastomosing greenstone belts with some evidence of early thrusting. The greenstone belts are cut by ovoid batholiths. The margins of the subprovince are characterized by long, relatively narrow, shear-zone-bounded panels of supracrustal rocks.

Mineral deposits of the subprovince include volcanogenic copper-gold and zinc-copper-silver deposits within volcanic units and Algoma-type iron formation at Atikokan. Mineralization associated with igneous rocks includes: pegmatite-related rare metal, uranium and molybdenum deposits, copper, copper-nickel, chromite and platinum group element deposits associated with mafic and ultramafic intrusions. Gold deposits related to late shear zones and contact strain aureoles developed around batholiths are found throughout the subprovince.

INTRODUCTION

The Wabigoon Subprovince (Figure 9.1) is a 900 km-long, 150 km-wide granite-greenstone subprovince in northwestern Superior Province. It comprises (Figure 9.2) metamorphosed volcanic and subordinate sedimentary rocks, ranging in age from about 3 to 2.71 billion years old, cut by *circa* 3 to 2.69 billion-year-old granitoid batholiths, gabbroic sills and stocks. The subprovince was the site of early studies defining classical terms such as Keewatin and Coutchiching supracrustal units (Lawson 1885, 1888, 1913). Geological investigations continued through the classical period into some of the early modern sedimentological investigations by Pettijohn (1935). Geologic and geochronologic research within the subprovince has been intense in the last few years (e.g., Trowell et al. 1980; Davis et al. 1988, 1989). The subprovince was defined by Goodwin (1970), but the boundaries of the subprovince used for this chapter are those shown by Card and Ciesielski (1986). The subprovince is one of the most extensively mapped regions in Ontario, supported by numerous U-Pb age determinations.

This chapter will emphasize stratigraphic relationships within greenstone belts and the relationship between batholithic complexes and supracrustal units. Greenstone belts of the Wabigoon Subprovince are almost uniformly of low

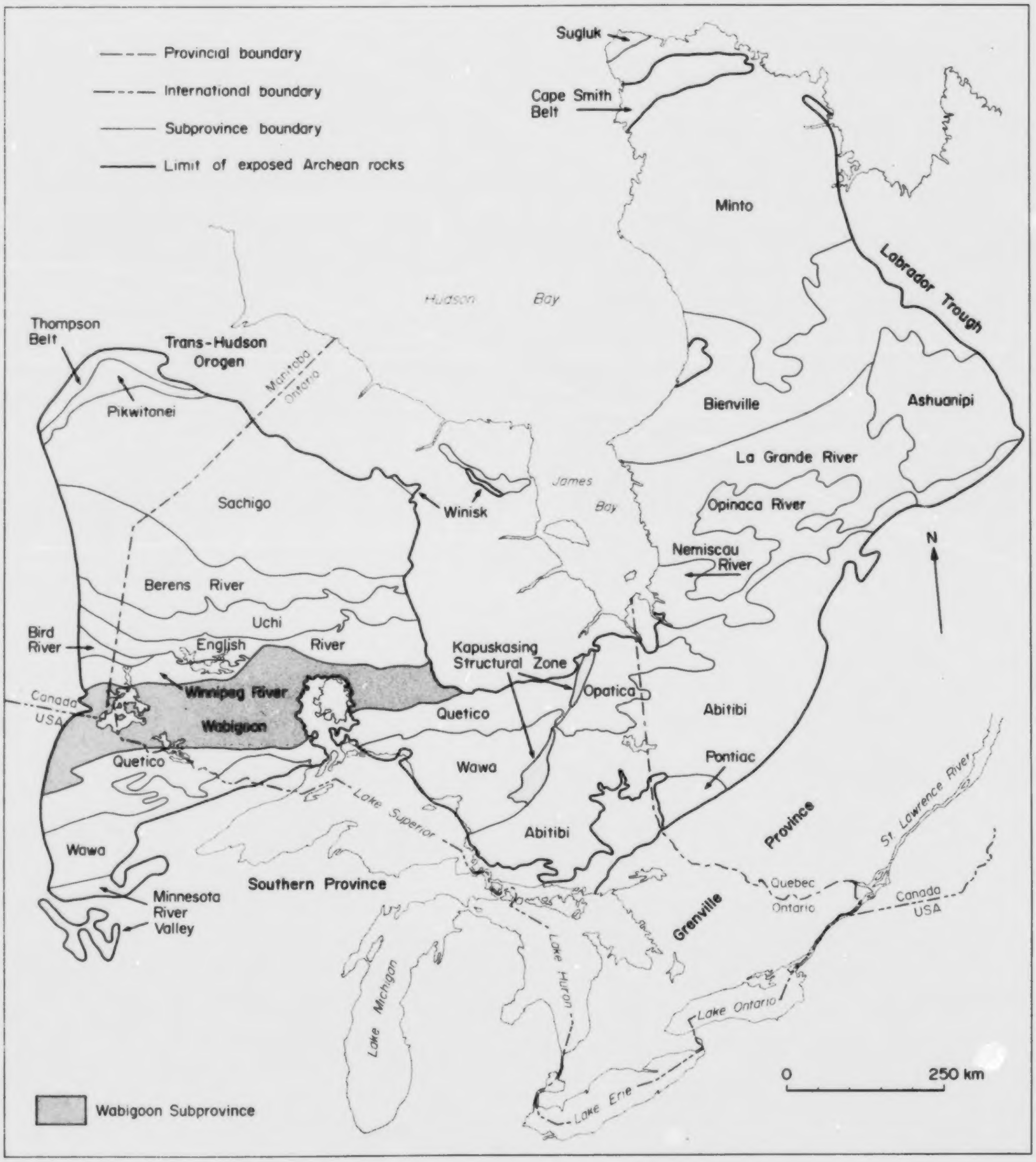

Provincial boundary
International boundary
Subprovince boundary
Limit of exposed Archean rocks
Sugluk
Cape Smith Belt
Minto
Labrador Trough
Hudson Bay
Thompson Belt
Trans-Hudson Orogen
Manitoba
Ontario
Pikwitonei
Bienville
Ashuanipi
La Grande River
James Bay
Opinaca River
Sachigo
Winisk
Nemiscau River
N
Berens River
Uchi
Bird River
English River
Kapuskasing Structural Zone
Winnipeg River
Canada
USA
Wabigoon
Opatica
Quetico
Abitibi
Quetico
Wawa
Pontiac
Lake Superior
Abitibi
Province
St. Lawrence River
Wawa
Southern Province
Quebec
Ontario
Grenville
Canada
USA
Minnesota River Valley
Lake Michigan
Lake Huron
Lake Ontario
0
250 km
Lake Erie
Wabigoon Subprovince

Figure 9.1. Location of the Wabigoon Subprovince within the Superior Province (*modified from* Card and Ciesielski 1986).

metamorphic grade; they are cut by elliptical batholithic complexes. Granitic areas of the subprovince are amongst the best known in the Superior Province. Geochemical and geochronologic evidence indicates the large elliptical internal batholiths are penecontemporaneous and consanguineous with the greenstones surrounding them. Granitic areas and greenstone belts are cut by posttectonic stocks.

The Wabigoon Subprovince is bordered to the northwest by the metaplutonic Winnipeg River Subprovince, to the northeast by the metasedimentary to migmatitic English River Subprovince and to the south by the metasedimentary Quetico Subprovince. The western boundary, beneath Phanerozic cover in Manitoba, is determined by the rocks of the 1.8 billion-year-old Trans–Hudson Orogen. The subprovince is cut by Proterozoic northwest-trending diabase dikes and sheets of the Nipigon Embayment.

The subprovince is divided into 3 parts based on geographic distribution of lithologic associations (*see* Figure 9.2): 1) the western Wabigoon region (WWR), underlain by supracrustal rocks and lobate intravolcanic granitoid batholiths; 2) the central Wabigoon region (CWR), underlain by ovoid gneissic domes and elliptical batholiths with screens and small belts of supracrustal rocks; and 3) the eastern Wabigoon region (EWR), underlain by supracrustal rocks and lobate intravolcanic granitoid batholiths. Relations between these 3 regions of the subprovince are not yet fully understood. Boundaries are ill-defined, with interfingering contacts: some greenstone sequences of the WWR extend into the CWR as amphibolite screens or trains of mafic inclusions within granitoid rocks. Small greenstone belts near the west side of Lake Nipigon may be correlated with the EWR, for example, in the Caribou Lake area, (Sutcliffe 1986) north of Lake Nipigon; therefore, these belts are considered part of the EWR.

The greatest amount of work has been done in the western Wabigoon region, therefore this region will be emphasized in this account; the eastern Wabigoon region will be discussed mainly by comparison with the western Wabigoon region; and the central Wabigoon region will be discussed insofar as it bears on understanding its relationship to the other 2 regions.

Brief summaries of the 3 parts of the subprovince follow, in order to set the scene for the descriptions of the major lithologic associations.

This chapter differs from the other chapters describing granite-greenstone subprovinces in that the description of the supracrustal rocks is on a lithostratigraphic basis rather than on a tectonostratigraphic basis as is done, for example, in the Uchi Subprovince. The reason for this is that the lithostratigraphic correlations have been emphasized in most of the mapping and modern structural studies are of a scattered nature . Trowell et al. (1980) subdivided the western region of the subprovince into a lower mafic sequence, a diverse intermediate to felsic sequence and an upper mafic sequence. Each sequence comprises several assemblages in the sense in which the term is used in the other chapters on granite-greenstone subprovinces of the Superior Province. Assemblages are shown on the tectonic assemblage map (*see* maps 2576 and 2577, map case) accompanying the volume. These sequences or groupings of assemblage types occur in scattered locations throughout the WWR. Each type of sequence (as defined by Trowell et al. 1980) possesses similar stratigraphy, geochemistry, and to some degree, similar absolute ages.

Western Wabigoon Region

The western Wabigoon region (WWR) is a series of interconnected greenstone belts surrounding large elliptical granitoid batholiths. The major greenstone belts are shown in Figure 9.2. Volcanic sequences comprise ultramafic (komatiitic), through mafic (tholeiitic, calc-alkalic, and minor alkalic and komatiitic) types, to felsic (mostly calc-alkalic) rocks. Sedimentary sequences are mostly clastic rocks of alluvial fan-fluvial, resedimented (turbidite) and rare platformal facies. Minor chemical metasedimentary rocks are predominantly oxide iron formation. Major granitoid batholiths are indicated on Figure 9.2. Numerous smaller posttectonic granitoid stocks intrude the greenstone belts. Mafic to ultramafic sills and stocks are marginal to batholiths or intrude the metavolcanic sequences.

Mafic metavolcanic units, commonly at the base of supracrustal sequences, have rarely been dated; the oldest unit is a 2775±1 million-year-old interflow tuff (Davis et al. 1988) in the Sturgeon Lake belt. Most felsic to intermediate volcanism within the WWR occurred in the interval 2745 to 2711 Ma, coeval with the early, marginal phases of the internal batholiths. These largely metavolcanic units are overlain by synorogenic metasedimentary units of a resedimented facies association or less commonly by alluvial fan-fluvial metasedimentary rocks. Deformation and syntectonic to posttectonic plutonism occurred in the interval 2711 to 2685 Ma.

Central Wabigoon Region

The central Wabigoon region (CWR) is dominated by large areas of granitoid and gneissic rocks. Major batholiths are shown on Figure 9.2. Based upon limited field data, the CWR consists of some of the oldest foliated and gneissic domical bodies within the Wabigoon Subprovince, consisting of tonalitic rocks cut and surrounded by younger massive and foliated granitic bodies (Sage et al. 1973) forming large-scale dome and basin structures (Edwards and Sutcliffe 1980). Granitic units are cut by metamorphosed mafic dike swarms at Steep Rock Lake (Wilks and Nisbet 1988) north of Armstrong (Sutcliffe 1986), and southwest of Armstrong (Cortis et al. 1988). Minor greenstone belts within the CWR are of 2 types: belts restricted to the CWR and screens and dismembered arms of greenstone belts of the adjoining regions. Scarce preliminary zircon data indicates some gneisses are older than 3075 million years, several units are about 2.75 billion years old, whereas others are 2708 million years old (Davis et al. 1988). Thurston and Davis (1985) suggested the CWR may represent a basement complex as is detailed in a subsequent section of this chapter. Cortis et al. (1988) have mapped an unconformity between tonalite cut by mafic dikes and the greenstones on the south margin of the Obonga Lake greenstone belt (Kustra 1966).

Eastern Wabigoon Region

The eastern Wabigoon region consists largely of isolated greenstone septa surrounded by granitoid units (*see* Figure 9.2). The domain is transected by the Proterozoic Nipigon Embayment. Greenstone belts can be traced, largely on geophysical evidence (*see* Maps 2584 to 2587, map case) from east of Lake Nipigon, beneath Proterozoic units, to the western side of the lake. Volcanic sequences are similar to those within the WWR, and lack only the alkalic metavolcanic rocks. Clastic metasedimentary rocks are much less abundant than in the WWR except along the northern and southern margins of the WWR, which are transitional to the English River and Quetico subprovinces.

Unlike the WWR, EWR metavolcanic units are clustered concentrically around volcanic "centres" (*see* Figures 9.2 and 9.3) at northern Marshall Lake; the Gzowski–Oboshkegan area; and the Elmhirst–Rickaby centre (Blackburn and Johns 1988). The granitoid batholiths and batholithic complexes are shown on Figures 9.2 and 9.3. As in WWR, numerous smaller granitoid, and gabbroic to dioritic stocks intrude the greenstone belts.

In contrast to the WWR, where most of the volcanism is confined to the interval 2745 to 2711 Ma, the limited geochronologic work undertaken in the region has shown widely varying ages for volcanism. An age of 2769^{+6}_{-5} Ma (Anglin et al. 1988) from a rhyolite in the Onaman River septum, between the Caribou–O'Sullivan and Elmhirst–Castlewood–Klotz lakes belts, is older than most felsic activity in the WWR. The Calm Lake–Lac des Mille Lacs greenstone belt (*see* Figure 9.2), between Lake Nipigon and Rainy Lake, provides a physical link between the greenstones of the EWR and the WWR. A felsic tuff from Calm Lake is 2722±1 million years old (Davis et al. 1989). However, pre-2.8 million-year-old model Pb ages from volcanic rocks of the EWR suggest that some units may be older than

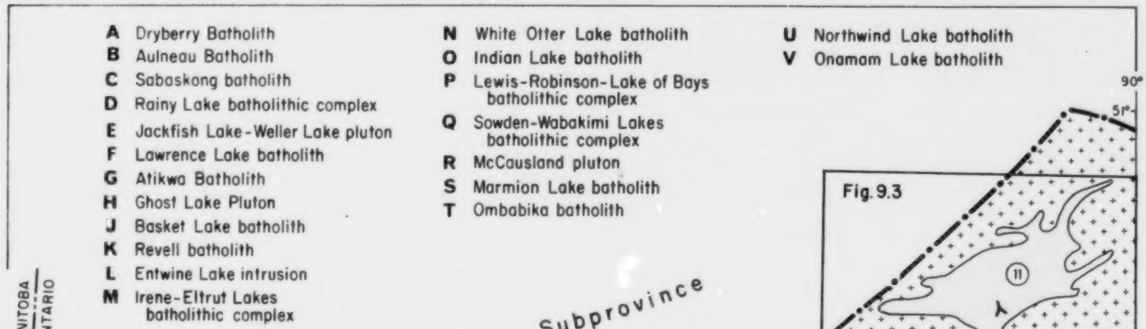

A Dryberry Batholith
B Aulneau Batholith
C Sabaskong batholith
D Rainy Lake batholithic complex
E Jackfish Lake-Weller Lake pluton
F Lawrence Lake batholith
G Atikwa Batholith
H Ghost Lake Pluton
J Basket Lake batholith
K Revell batholith
L Entwine Lake intrusion
M Irene-Eltrut Lakes batholithic complex
N White Otter Lake batholith
O Indian Lake batholith
P Lewis-Robinson-Lake of Bays batholithic complex
Q Sowden-Wabakimi Lakes batholithic complex
R McCausland pluton
S Marmion Lake batholith
T Ombabika batholith
U Northwind Lake batholith
V Onamam Lake batholith
MANITOBA
ONTARIO
Winnipeg River Subprovince
Quetico Subprovince
WESTERN
CENTRAL
EASTERN
Fig.9.3
Fig.9.12
Fig.9.9
Fig.9.10
Fig. 9.8
Fig.9.11
Fig.9.13
Sioux Lookout
Savant Lake
Vermilion Bay
Kenora
Dryden
Sioux Narrows
Ignace
Atikokan
Mine Centre
Fort Frances
Minnesota
95° + 50°
+ 49° 95°
90° 51°
50°
49°
90°

Figure 9.2. General geology of the Wabigoon Subprovince in Ontario, showing its extent, major supracrustal belts and structural features, and areas detailed in subsequent figures.

those of western Wabigoon (R. Thorpe, Geological Survey of Canada, personal communication).

Stratigraphic Nomenclature

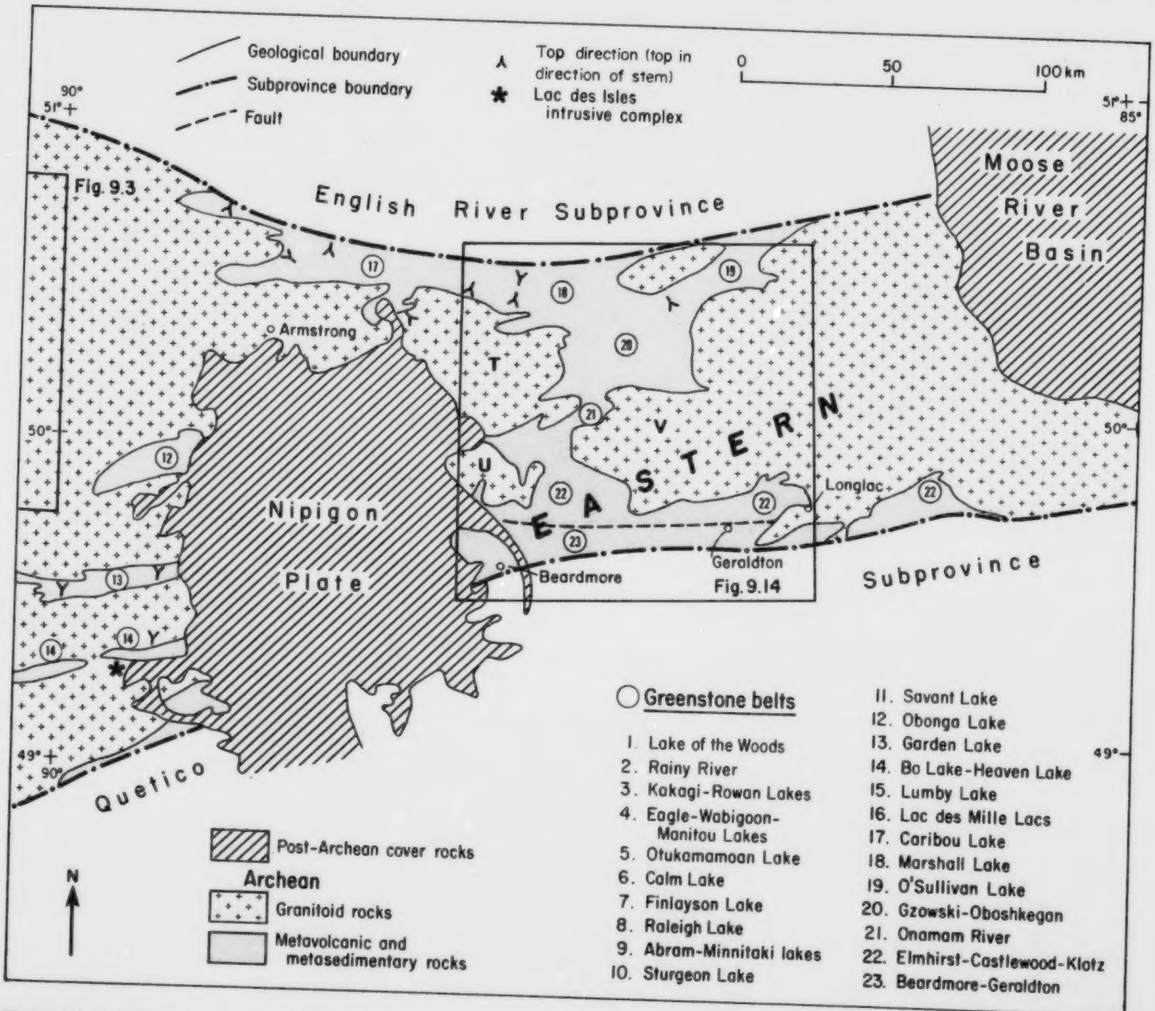
Unit nomenclature in Wabigoon Subprovince dates back to Lawson (1885, 1888) who introduced the current terms Keewatin and Coutchiching. In this chapter, nomenclature will conform with the North American Code of Stratigraphic Nomenclature (NACSN 1983). Comparison with former usage is presented in Tables 9.1 and 9.2, and Figures 9.4, 9.5, 9.6 and 9.7. The description of units in the Wabigoon Subprovince is lithostratigraphic. In a later section, these lithostratigraphic units are related to tectonic assemblages in common with the other Superior Province chapters in this volume.

LITHOLOGIC DESCRIPTIONS

Field relationships, petrography and geochemical descriptions of rock units are presented in this section. The regional-scale relationships amongst greenstone belts are seen on Figure 9.2. More detailed maps of individual greenstone belts are presented as Figures 9.3, 9.8, 9.9, 9.10, 9.11, 9.12, 9.13 and 9.14, which should be consulted in conjunction with the lithologic descriptions. The descriptions use revised stratigraphic nomenclature in many instances. Tables 9.1 and 9.2, and Figures 9.4, 9.5, 9.6 and 9.7 trace the history of stratigraphic nomenclature for major supracrustal units of the Wabigoon Subprovince.

Metavolcanic Rocks and Subvolcanic Intrusions

Greenstone belts of the Wabigoon Subprovince are composed of 60 to 80% metavolcanic rocks with the remainder



Geological boundary
Subprovince boundary
Fault
Top direction (top in direction of stem)
Lac des Isles intrusive complex
0
50
100 km
90°
51°
51°
85°
50°
50°
49°
49°
90°
Fig. 9.3
English River Subprovince
Moose River Basin
Armstrong
Nipigon Plate
EASTERN
Longlac
Geraldton
Beardmore
Fig. 9.14
Subprovince
Quetico
T
V
U
Y
11
12
13
14
17
18
19
20
21
22
23
Post-Archean cover rocks
Archean
Granitoid rocks
Metavolcanic and metasedimentary rocks
N
Greenstone belts
1. Lake of the Woods
2. Rainy River
3. Kakagi-Rowan Lakes
4. Eagle-Wabigoon-Manitou Lakes
5. Otukamamoan Lake
6. Calm Lake
7. Finlayson Lake
8. Raleigh Lake
9. Abram-Minnitaki lakes
10. Sturgeon Lake
11. Savant Lake
12. Obonga Lake
13. Garden Lake
14. Bo Lake-Heaven Lake
15. Lumby Lake
16. Lac des Mille Lacs
17. Caribou Lake
18. Marshall Lake
19. O'Sullivan Lake
20. Gzowski-Oboshkegan
21. Onamam River
22. Elmhirst-Castlewood-Klotz
23. Beardmore-Geraldton

Figure 9.2. General geology of the Wabigoon Subprovince in Ontario, showing its extent, major supracrustal belts and structural features, and areas detailed in subsequent figures.

being of metasedimentary origin. Of the metavolcanic rocks and coeval subvolcanic equivalents, about 70% are mafic tholeiites, and the remainder are calc-alkalic intermediate to felsic rocks with only minor alkalic and komatiitic rocks. Numerous workers (Goodwin 1965, 1970; Wilson et al. 1974; Trowell et al. 1980; Blackburn et al. 1985; Trowell and Johns 1986; Morrice 1988, 1989; Ayer 1987, 1988a; Ayer and Buck 1989) have demonstrated that the Wabigoon Subprovince metavolcanic rocks can be grouped into 2 categories: 1) basal, laterally extensive, mafic metavolcanic rocks; subordinate mafic metavolcanic units occupy stratigraphically higher positions; and 2) less thick, laterally limited, diverse assemblages of intermediate to felsic volcanic rocks that commonly overlie mafic metavolcanic assemblages. Absolute age data (*see* Discussion *in* Metasedimentary Rocks) and mapping (*see* Structural Geology) have demonstrated the presence of thrust-based duplication of strata, therefore the classical, simplistic models for greenstone belts involving central synclinoria are of limited use. Hence, we attempt to describe the supracrustal rocks in terms of small-scale lithostratigraphic and tectonically based units.

MAFIC METAVOLCANIC SEQUENCES

The mafic volcanic sequences are up to 11 km thick, monotonous successions of pillowed and massive basaltic flows with minor interflow units. Within WWR, these successions occupy the 2 distinct stratigraphic positions discussed above (*see* Metavolcanic Rocks and Subvolcanic Intrusions). However, in EWR, this relationship is less well documented.

Lower Mafic Sequences of Western Wabigoon

Lower mafic sequences of WWR are listed in Table 9.3. The apparent thickness of lower mafic sequences may be a function of tectonic repetition (*see* Structural Geology). Neither structural nor stratigraphically based geochemical studies have been done to identify tectonic repetition on a consistent regional basis. The sequences consist of pillowed and massive mafic flows with minor fragmental rocks. Bases of lower mafic sequences are invariably obscured by intrusion of batholiths, by faulting, or are concealed within the cores of anticlinal structures (*see* Table 9.3). The sequences typically face away from batholiths, for example, the Irene–Eltrut Lakes batholithic complex relative to the Wapageisi group (*see* Figure 9.8).

The transition from lower mafic sequences into overlying sequences may be conformable to unconformable. Examples of conformable transitions from a tholeiitic lower mafic sequence into overlying calc-alkalic diverse volcanic sequences are: the top of the Bigstone Bay group (Ayer et al.

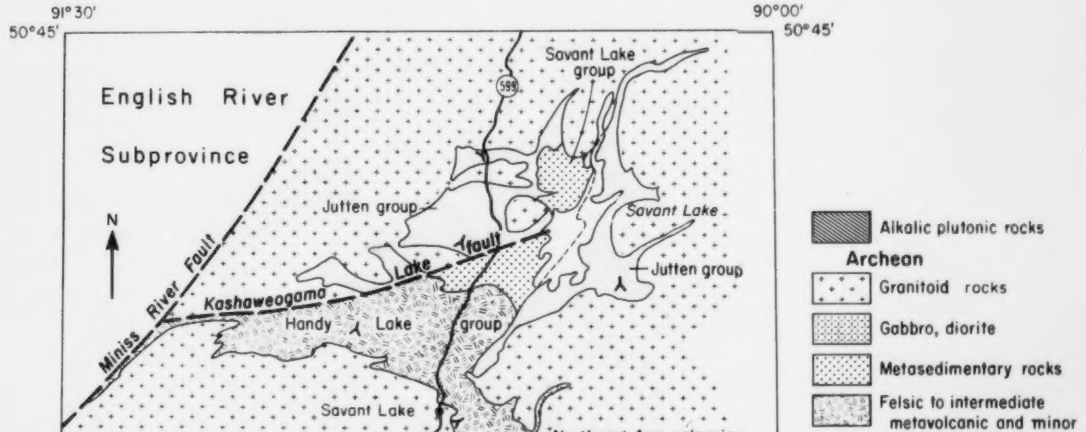

91°30'
50°45'
90°00'
50°45'
English River Subprovince
N
Savant Lake group
599
Jutten group
Savant Lake
Jutten group
Minniss River Fault
Kashaweogama Lake fault
Handy Lake group
Savant Lake
Lewis-Robinson-Lake of Bays batholithic complex
Northeast Arm volcanics
Sturgeon Lake fault
North Sturgeon Lake volcanics
Lake
Central Sturgeon Lake volcanics
Sturgeon Narrows Alkalic complex
Quest Lake sediments
Sturgeon
642
South Sturgeon Lake volcanics
Beidelman Bay pluton
0 10 20km
49°45'
91°30'
49°45'
90°00'
Alkalic plutonic rocks
Archean
Granitoid rocks
Gabbro, diorite
Metasedimentary rocks
Felsic to intermediate metavolcanic and minor metasedimentary rocks
Mafic metavolcanic rocks
Geological boundary (observed, assumed)
Fault
Top direction (top in direction of stem)
Highway

Figure 9.3. Detailed geology, Sturgeon Lake greenstone belt and Savant Lake greenstone belt.

Table 9.1. Stratigraphic nomenclature, Lake of the Woods greenstone belt.

Lawson (1885)	Thomson (1937)	Goodwin (1970)	Wilson and Morris (1977)	This Chapter	
				Posttectonic granitoids	
Late Granites and Felsites	Algoman Granitoids Timiskaming series			Electrum Lake supergroup	Crowduck Lake group White Partridge Bay group
				Syntectonic granitoid	
Keewatin Series	Keewatin Series	Upper Keewatin Sequence Lower Keewatin Sequence	Upper Diverse Group Upper Felsic Group	Upper Keewatin Supergroup	Clearwater Bay group Indian Bay group Andrew Bay group Monument Bay group Royal Island group Warclub group Long Bay group Windigo Islands group
			Middle Mafic Group Lower Mafic Group	Lower Keewatin Supergroup	Deception Bay group Bigstone Bay group Cedar Island group Barrier Island group Snake Bay group
Laurentian Gneisses				Winnipeg River Subprovince orthogneiss	

Table 9.2. Stratigraphic nomenclature, Fort Frances–Mine Centre greenstone belt.

Lawson (1913)	Grout (1925)	Harris (1974)	Wood (1980)	Poulsen et al. (1980)
Algoman granitoids	Algoman granitoids	Algoman granitoids	Algoman granitoids	Algoman granitoids
Intrusive contact	Intrusive contact	Intrusive contact	Intrusive contact	Intrusive contact
Seine metasediments	Couchiching metasediments Seine metasediments	Laurentian gneisses	Coutchiching–Seine metasediments (facies equivalent)	Seine metasediments
Unconformity	Unconformity	Intrusive contact	Unconformity	Unconformity
Laurentian gneisses	Laurentian gneisses	Keewatin metavolcanics–Coutchiching metasediments (intercalated)	Laurentian gneisses	Laurentian gneisses
Intrusive contact	Intrusive contact		Intrusive contact	Intrusive contact
Keewatin metavolcanics Coutchiching metasediments	Keewatin metavolcanics		Keewatin metavolcanics	Coutchiching metasediments Keewatin metavolcanics

1987; *see* Figures 9.2 and 9.9), the Rowan Lake Volcanics (Figure 9.10; Melling et al. 1988) and the Blanchard Lake group (Blackburn 1982; *see* Figure 9.8). Other conformable contacts are marked by epidote-cemented regolithic breccia and pahoehoe toes (Snake Bay group, Johns 1985; *see* Figure 9.10), or thin discontinuous sedimentary horizons (Katimiagamak group, Edwards 1980; *see* Figure 9.10). Some contacts are abrupt, locally marked by minor epiclastic metasedimentary rocks, others are gradational, with interdigitation of mafic flows and felsic intermediate pyroclastic rocks.

In more detail, lower mafic sequences consist of flow units from 5 to 20 m thick, predominantly simple, and uniformly pillowed or massive throughout (Figure 9.15). Compound flows, grading from massive into pillowed parts are subordinate. Vesicles, varioles, pillow breccia, autoclastic flow breccia, and hyaloclastite are generally limited to stratigraphically higher levels within the sequences (Figure 9.16). The limited vesiculation and absence of flow breccias in the lower part of the successions implies eruption and deposition in deep water on gentle slopes or platforms. The localization of breccias and large vesicles in the upper parts suggests the lower mafic sequences commonly shallow upwards (cf., Wilson and Morrice 1977). Shallower water is also indicated by the primary structures and eruptive mechanisms of the immediately overlying sequences (e.g., abundant pyroclastic rocks and highly vesiculated flows of the diverse volcanic sequences).

Laterally extensive plagioclase-megacrystic basalt forms pillowed and massive flows, and sills, forming marker horizons in the lower mafic sequences (Figure 9.17; *see* Table 9.3). The marker horizons occur near the top of the Bigstone Bay group (*see* Figure 9.9), the Snake Bay group (*see* Figure 9.9), the Brooks Lake volcanics (*see* Figure 9.10), and the South Sturgeon Lake volcanics (*see* Figure 9.3). They also form laterally extensive horizons in lower parts of the Snake Bay group, and the Wapageisi group.

Phinney et al. (1988) have reasoned that plagioclase megacrystic units in mafic flow sequences are related to the development of ponded crustal magma chambers. They envisage a multistage petrogenesis involving olivine and/or orthopyroxene fractionation followed by ascent of the fractionated, less dense melt to a low-pressure chamber. The melts would crystallize as anorthositic complexes and periodically expel megacrystic flows and subvolcanic sills.

Lower mafic sequences are overwhelmingly of tholeiitic basalt composition. An iron-enrichment trend from magnesium tholeiite to iron tholeiite is evident in some successions, for example, the Snake Bay group (Wilson and Morrice 1977), the Wapageisi group (Blackburn 1982) and the Bigstone Bay group (Ayer et al. 1987). However, an iron-enrichment trend is not always evident (e.g., the Jutten group; Trowell 1986a). The lower parts of many of the lower mafic sequences (including the Jutten group and the Wapageisi group) contain a minor proportion of magnesium-rich basaltic komatiite to ultramafic komatiite volcanic rocks (Blackburn 1982; Trowell 1986a).

The lower mafic sequences exhibiting iron-enrichment trends are characterized by upward depletion in magnesium, chromium, nickel, and enrichment upward in K_2O, FeO[1] and Ti. Chondrite-normalized rare earth diagrams show flat to slightly depleted light rare earth elements (LREE) at about 10 times chondritic levels in the lower parts of the units ($[La/Yb]n = 0.7$-1.0) and flat patterns at about 20 times chondritic levels in upper parts of units ($[La/Yb]_N = 1.0$-1.5), Ayer and Buck 1989; Ayer, in press). The lack of Eu anomalies indicates that plagioclase fractionation was not likely involved in the petrogenesis of the lower mafic sequences.

Although geochemically comparable with modern ocean-floor basalts, the lower mafic sequences bear most similarity to low K_2O island-arc tholeiites (Gill 1979). The generally flat to slightly LREE-depleted patterns of the stratigraphically lower sections of the sequence suggest partial melting of an undepleted to slightly depleted upper mantle (cf., Hanson 1980) and fractionation of olivine and plagioclase. However, the low magnesium, chromium and nickel values for the upper iron-rich sections implies fractionation of olivine or remelting of magnesium-rich basalts in crustal magma chambers.

Mafic to ultramafic intrusions occur in all of the lower mafic sequences and are a major component of the Bigstone Bay group, the Katimiagamak group, and the Rowan Lake volcanics. These intrusions range from homogeneous gabbro sills and dikes to differentiated sills. Differentiated sills within the Bigstone Bay group are up to 2 km thick and 12 km long. They typically display one or more cycles grading upwards from peridotite to pyroxenite into gabbro or leucogabbro (Ayer et al. 1987). Contacts with the metavolcanic rocks range from crosscutting to concordant (Figure 9.18). We consider the mafic intrusions to be synvolcanic on petrologic and geochronologic grounds (e.g., Davis and Edwards 1982). Indeed, some may be large-ponded flows in which slow cooling has resulted in a coarse grain size. This interpretation is supported by local gradation of coarse sills into fine-grained brecciated flows. In the Hades Islands (*see* Figure 9.9), the chilled base of a 100 m-thick differentiated sill conforms to the undulating bulbous shapes of the underlying pillowed flow (Ayer et al. 1987; Figure 9.19).

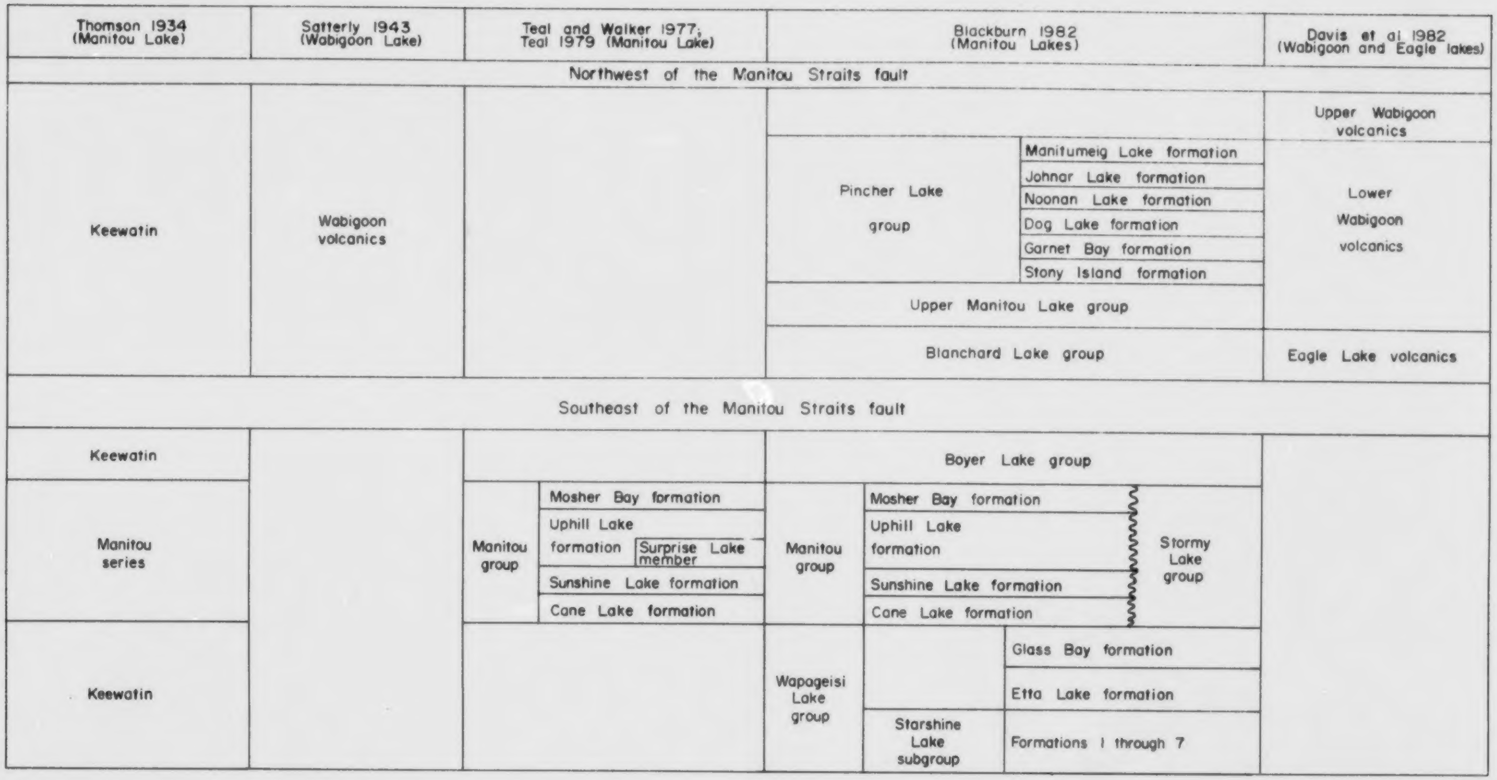
Minor interflow sedimentary units occur at various stratigraphic levels in lower mafic sequences, for example, the Etta Lake formation near the top of the Wapageisi group (Blackburn 1982) and chert-magnetite-banded iron formation near the top of the Northern volcanic belt (*see* Figure 9.12) and near the base of the Jutten group (*see* Figure 9.3). A thin unit of interbedded sandstone and heterolithic conglomerate, designated the Jutten sedimentary sequence (Sanborn-Barrie 1989), occurs within mafic flows of the southeastern part of the Jutten group (Trowell 1986a). Recent investigation of this unit has suggested that it may represent a medial- to distal-fan equivalent of a platform association of granitoid provenance (Cortis et al. 1988; Sanborn-Barrie 1989).

---

[1] *Total iron expressed as FeO.*

Burwash 1934	Davies and Watowitch 1956	This volume (NE of Pipestone-Cameron deformation zone)				Burwash 1934	Davis and Edwards 1982	This volume (SW of Pipestone-Cameron deformation zone)		
Upper Keewatin	Warclub Sediments	Warclub group	Berry River formation		Pipestone – Cameron deformation zone	Upper Keewatin				
		Hill Lake volcanics = Otterskin Lake volcanics		Upper Mafic group			Kakagi Lake Group			Upper Mafic group
		Brooks Lake volcanics								
		Cameron Lake volcanics	Upper Member	Upper Diverse group				Kakagi Lake group	Stephen Lake formation	Upper Diverse group
									Cedartree Lake formation	
			Lower Member						Emm Bay formation	
									East Kakagi Lake formation	
		Dogpaw Lake volcanics							South Kakagi Lake formation	
Lower Keewatin	Populus Volcanics	Rowan Lake volcanics = Populus Lake volcanics		Lower Mafic group		Lower Keewatin	Snake Bay Formation = Katimiagamak Formation	Snake Bay group = Katimiagamak group		Lower Mafic group
Coutchiching (Warclub Sediments)		Point Bay group				Coutchiching (Warclub Sediments)				

Figure 9.4. Stratigraphic nomenclature, Kakagi–Rowan Lakes greenstone belt. Rock types are listed from youngest to oldest, in descending order.



Thomson 1934 (Manitou Lake)	Satterly 1943 (Wabigoon Lake)	Teal and Walker 1977; Teal 1979 (Manitou Lake)	Blackburn 1982 (Manitou Lakes)		Davis et al 1982 (Wabigoon and Eagle lakes)
Northwest of the Manitou Straits fault					
Keewatin	Wabigoon volcanics				Upper Wabigoon volcanics
			Pincher Lake group	Manitumeig Lake formation Johnar Lake formation Noonan Lake formation Dog Lake formation Garnet Bay formation Stony Island formation	Lower Wabigoon volcanics
			Upper Manitou Lake group		
			Blanchard Lake group		Eagle Lake volcanics
Southeast of the Manitou Straits fault					
Keewatin			Boyer Lake group		
Manitou series		Manitou group Mosher Bay formation Uphill Lake formation Surprise Lake member Sunshine Lake formation Cane Lake formation	Manitou group Mosher Bay formation Uphill Lake formation Sunshine Lake formation Cane Lake formation	Stormy Lake group	
Keewatin			Wapageisi Lake group	Glass Bay formation Etta Lake formation	
			Starshine Lake subgroup	Formations 1 through 7	

Figure 9.5. Stratigraphic nomenclature, Eagle–Wabigoon–Manitou Lakes greenstone belt.

<table>
<tr><th colspan="2">Pettijohn, 1935, 1943 (Abram Lake)</th><th>Pettijohn, 1936 (Minnitaki Lake)</th><th>Satterly 1943 (Wabigoon Lake)</th><th colspan="2">Turner and Walker 1973</th><th colspan="3">Beakhouse 1988</th><th>Present volume</th></tr>
<tr><td colspan="10">North of the Wabigoon fault</td></tr>
<tr><td colspan="2"></td><td>Older Keewatin Lavas
or
Pre-Abram volcanics</td><td></td><td colspan="2">Southern volcanic belt</td><td colspan="3">Southern volcanic belt</td><td>Southern volcanic belt</td></tr>
<tr><td colspan="10">South of the Wabigoon fault</td></tr>
<tr><td colspan="2" rowspan="3"></td><td rowspan="3">Abram series</td><td>Zealand sediments</td><td colspan="2" rowspan="3">Minnitaki Group (Southern sedimentary belt)</td><td colspan="3" rowspan="3">Southern sedimentary belt</td><td rowspan="3">Minnitaki group</td></tr>
<tr><td>Thunder River volcanics</td></tr>
<tr><td>Thunder Lake sediments</td></tr>
<tr><td>Keewatin</td><td>Older volcanics</td><td>Neepawa volcanics</td><td>Brownridge volcanics</td><td colspan="2">Central volcanic belt</td><td colspan="3">Central volcanic belt</td><td>Neepawa group</td></tr>
<tr><td rowspan="3">Abram series</td><td rowspan="2">Daredevil Formation</td><td rowspan="6"></td><td rowspan="3">Brownridge sediments</td><td rowspan="3">Abram Group (Northern sedimentary belt)</td><td>Little Vermilion Formation</td><td rowspan="5">Northern sedimentary belt</td><td rowspan="3">Abram Group</td><td>Little Vermilion Formation</td><td rowspan="3">Abram group</td></tr>
<tr><td>Daredevil Formation</td><td>Daredevil Formation</td></tr>
<tr><td>Ament Bay arkosite</td><td>Ament Bay Formation</td><td>Ament Bay Formation</td></tr>
<tr><td rowspan="3">Keewatin</td><td rowspan="2">Patara volcanogenic metasediments</td><td rowspan="3"></td><td rowspan="3">Northern volcanic belt</td><td>(volcanics)</td><td colspan="2" rowspan="2">Patara sediments and volcanics</td><td rowspan="2">Patara sediments</td></tr>
<tr><td>Patara sediments</td></tr>
<tr><td>Older volcanics</td><td>(volcanics)</td><td colspan="3">Northern volcanic belt</td><td>Northern volcanic belt</td></tr>
</table>

Figure 9.6. Stratigraphic nomenclature, Abram Lake greenston belt and Minnitaki Lake greenstone belt.

Lower mafic sequences range from 2775±1 Ma for the North Sturgeon Lake volcanics (*see* Figure 9.3), to 2732±1 Ma for the Katimiagamak group, based on synvolcanic gabbro sills. The lowermost units in greenstone belts need not represent the oldest rocks because of structural superposition. For example, the Wapageisi group, Jutten group and the Northern volcanic belt are all unconformably overlain by late *circa* 2700 million-year-old metasedimentary rocks, which are in turn structurally overlain by older *circa* 2750 to 2720 million-year-old mafic and diverse metavolcanic sequences. Another example is the greater than 2722 million-year-old Boyer Lake group, which may be correlative with other lower mafic metavolcanic sequences, although it occupies an upper structural position.

Upper Mafic Sequences of Western Wabigoon

In WWR, upper mafic sequences (Table 9.4) are characterized by monotonous pillowed and massive mafic flows. These sequences are generally thinner and less laterally extensive than the lower mafic sequences. They occur within, or above, the diverse volcanic sequences and may have intercalations of intermediate to felsic volcanic and sedimentary rocks.

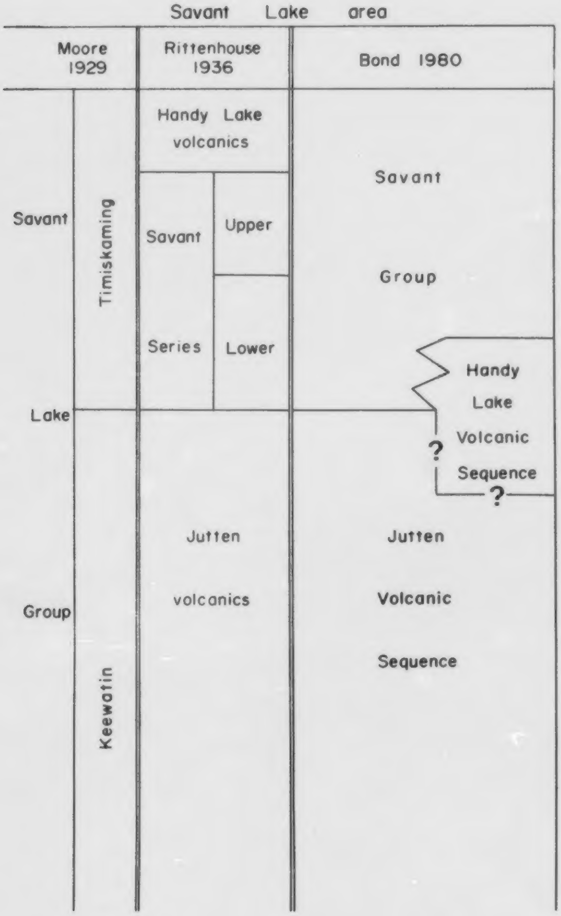


Figure 9.7. Stratigraphic nomenclature, Savant Lake greenstone belt.

Contacts with the diverse volcanic sequences are either faulted or conformable. The Upper Wabigoon volcanics (*see* Figure 9.8) are the thickest (approximately 6 km) and most extensive homoclinal upper mafic sequence. Upper mafic sequences occur in 2 stratigraphic positions relative to diverse volcanic sequences: 1) the Indian Bay group, Black Lake formation, Upper Wabigoon volcanics, and the Central Sturgeon Lake volcanics are near the top of diverse volcanic sequences; and 2) the Cochrane Island formation and Cliff Island formation occur at lower stratigraphic levels. Within upper mafic sequences, tholeiitic basalts are pillowed to massive; ultramafic komatiites are typically massive to polyhedral jointed, rarely pillowed and small vesicles occur locally (Ayer, in press). In contrast, vesicles of nearby occurrences of upper diverse sequences tend to be larger, suggesting that higher viscosity of the upper mafic sequences is the major control on vesicle size and abundance, rather than deep water effusion.

Varioles are reported in many of the sequences at Lake of the Woods, both within pillowed flows and capping unusual massive mafic flows with spinifex-like, coarse-grained, skeletal pyroxenes (Ayer and Buck 1989). These unusual massive flows and their close association with variolitic flows (Ayer and Buck 1989), indicate crystallization from a supercooled mantle-derived magma, similar to the Abitibi Subprovince examples (Fowler et al. 1987).

Metavolcanic rocks of the upper mafic sequence are quite variable in lithogeochemistry. Komatiitic basalt and ultramafic komatiite are common but subordinate to magnesium-rich tholeiites in the Cochrane Island and Cliff Island formations. The Upper Wabigoon volcanics and the Indian Bay group are iron tholeiitic basalts. The Cochrane Island formation, Cliff Island formation, and the Black Lake formation are similar, with slight LREE depletion or LREE enrichment and $[La/Yb]_N = 0.4$ to 1.4, at 7 to 12 times chondritic levels (Ayer and Buck 1989; Ayer, in press). In northern Lake of the Woods, the Indian Bay group tholeiites have slightly LREE-enriched patterns with $[La/Yb]_N = 0.8$ to 0.4, at 7 to 50 times chondritic levels (Ayer 1988a). They are therefore more fractionated than, and are transitional with, the mafic to intermediate rock types of the upper diverse volcanic sequence rather than being related to the more primitive upper mafic sequences in southern Lake of the Woods.

Mafic and ultramafic intrusions form gabbroic and less common peridotitic to gabbroic sills and dikes in many of the upper mafic sequences. Metasedimentary horizons and intermediate pyroclastic rocks occur in the Cliff Island formation, the Black Lake formation and the Central Sturgeon Lake volcanics. Metasedimentary rocks range from epiclastic units to chert and banded iron formation. Pyroclastic rocks are mostly resedimented and range from bedded tuffs to heterolithic tuff breccias

Mafic Sequences of Eastern Wabigoon

East of Lake Nipigon, the distinction between lower mafic sequences and mafic metavolcanic sequences lying within or above diverse volcanic rocks has not been made. Only locally has it been demonstrated that mafic flow sequences underlie felsic to intermediate, predominantly pyroclastic, sequences (Amukun 1979; Blackburn and Johns 1988). In general, the stratigraphic position of mafic flow sequences can be inferred only in a few instances.

Only the Caribou–O'Sullivan Lakes greenstone belts (*see* Figure 9.2) contain extensive mafic flows. They surround the Marshall Lake volcanic centre, and the Gzowski–Oboshkegan volcanic centre. Amukun (1979) has described basal mafic flows conformably overlain by felsic to intermediate pyroclastic rocks. The mafic volcanic rocks consist of subequal amounts of pillowed and massive tholeiitic basalt flows. Aphyric flows predominate, with locally developed plagioclase porphyritic and amygdaloidal flows.

Spinifex-textured komatiites are an integral part of the tholeiitic mafic volcanic succession in the northwestern part of the Caribou–O'Sullivan Lakes greenstone belt (Sutcliffe 1986). An U-Pb age of at least 3075 Ma was obtained from tonalite gneiss that intrudes the metavolcanic succession

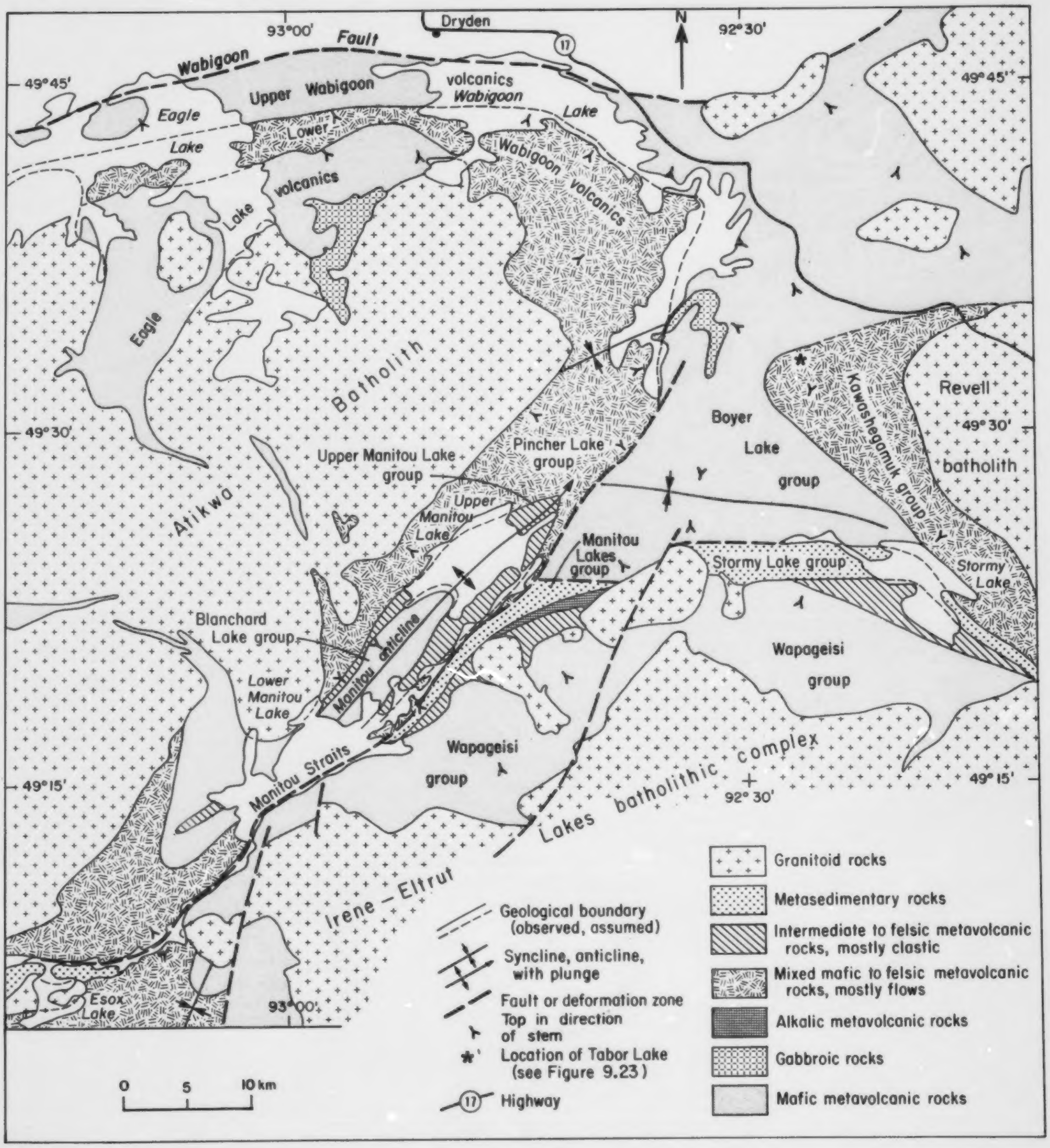

Dryden
93°00'
Fault
Wabigoon
N
92°30'
17
49°45'
Upper Wabigoon
volcanics
Wabigoon
Lake
Eagle
Lake
Lower
volcanics
Lake
Wabigoon
volcanics
Eagle
Batholith
Atikwa
49°30'
Upper Manitou Lake group
Pincher Lake group
Upper Manitou Lake
Boyer
Lake
group
Kawashegamuk group
Revell
batholith
Manitou Lakes group
Stormy Lake group
Stormy Lake
Blanchard Lake group
Lower Manitou Lake
Manitou anticline
Wapageisi group
Wapageisi group
Manitou Straits
complex
Lakes batholithic
Irene – Eltrut
49°15'
92°30'
Esox Lake
93°00'
0
5
10 km
Geological boundary (observed, assumed)
Syncline, anticline, with plunge
Fault or deformation zone
Top in direction of stem
Location of Tabor Lake (see Figure 9.23)
Highway
Granitoid rocks
Metasedimentary rocks
Intermediate to felsic metavolcanic rocks, mostly clastic
Mixed mafic to felsic metavolcanic rocks, mostly flows
Alkalic metavolcanic rocks
Gabbroic rocks
Mafic metavolcanic rocks

Figure 9.8. Detailed geology, Eagle–Wabigoon–Manitou Lakes greenstone belt.

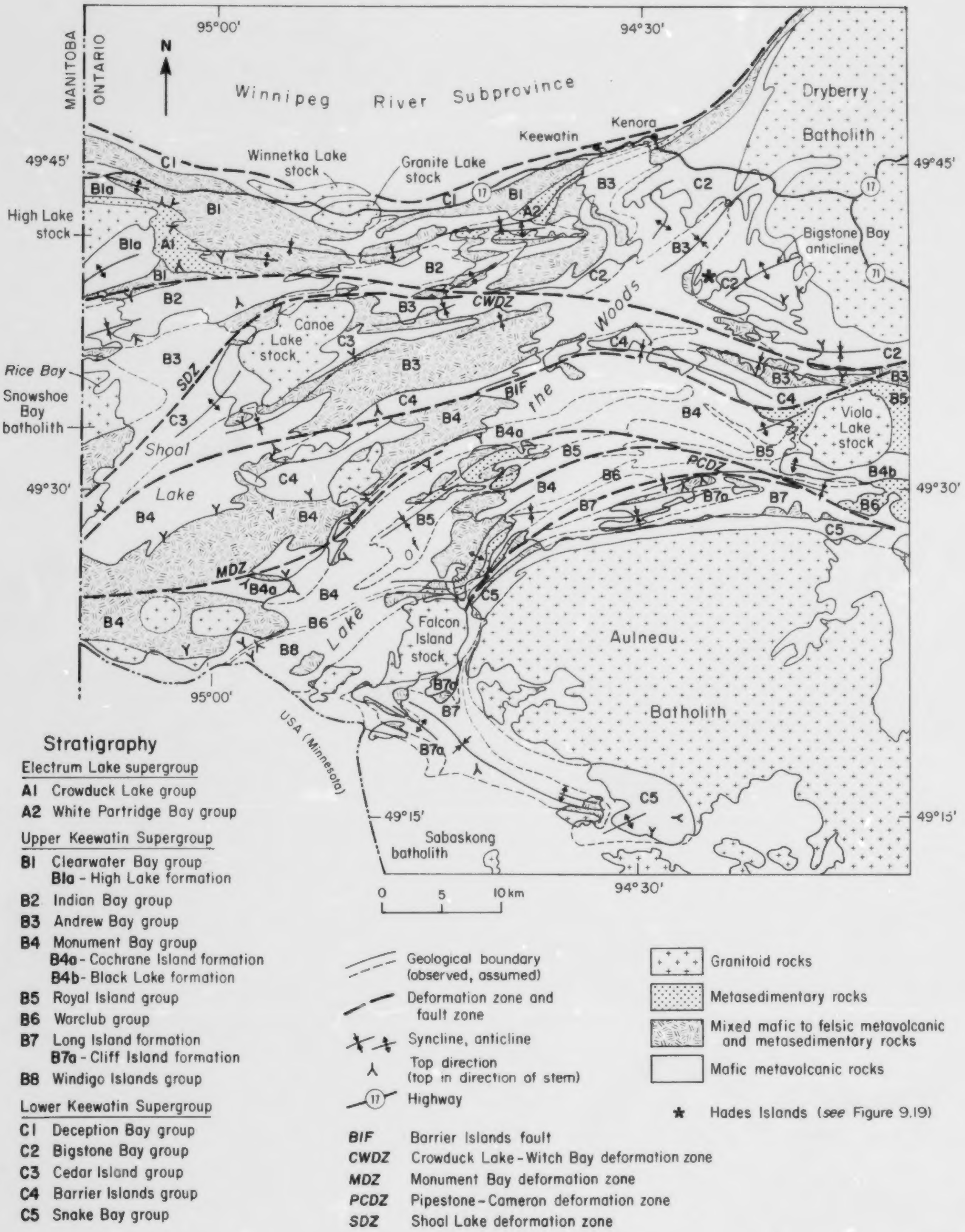

MANITOBA
ONTARIO
N
95°00'
94°30'
Winnipeg River Subprovince
Dryberry Batholith
Keewatin
Kenora
49°45'
Winnetka Lake stock
Granite Lake stock
High Lake stock
Rice Bay
Snowshoe Bay batholith
Canoe Lake stock
Bigstone Bay anticline
Viola Lake stock
Shoal Lake
Lake of the Woods
49°30'
Falcon Island stock
Aulneau Batholith
USA (Minnesota)
49°15'
Sabaskong batholith
CWDZ
SDZ
BIF
MDZ
PCDZ
A1
A2
B1
B1a
B2
B3
B4
B4a
B4b
B5
B6
B7
B7a
B8
C1
C2
C3
C4
C5
17
71
0
5
10 km
Stratigraphy
Electrum Lake supergroup
A1 Crowduck Lake group
A2 White Partridge Bay group
Upper Keewatin Supergroup
B1 Clearwater Bay group
B1a - High Lake formation
B2 Indian Bay group
B3 Andrew Bay group
B4 Monument Bay group
B4a - Cochrane Island formation
B4b - Black Lake formation
B5 Royal Island group
B6 Warclub group
B7 Long Island formation
B7a - Cliff Island formation
B8 Windigo Islands group
Lower Keewatin Supergroup
C1 Deception Bay group
C2 Bigstone Bay group
C3 Cedar Island group
C4 Barrier Islands group
C5 Snake Bay group
Geological boundary (observed, assumed)
Deformation zone and fault zone
Syncline, anticline
Top direction (top in direction of stem)
Highway
BIF Barrier Islands fault
CWDZ Crowduck Lake-Witch Bay deformation zone
MDZ Monument Bay deformation zone
PCDZ Pipestone-Cameron deformation zone
SDZ Shoal Lake deformation zone
Granitoid rocks
Metasedimentary rocks
Mixed mafic to felsic metavolcanic and metasedimentary rocks
Mafic metavolcanic rocks
★ Hades Islands (see Figure 9.19)

Figure 9.9. Detailed geology, Lake of the Woods greenstone belt.

(Davis et al. 1988). Spinifex-textured komatiite flows and ultramafic pyroclastic rocks have also been recognized in: the Steep Rock Group (*see* Figure 9.11), the Lumby Lake greenstone belt, and the Bo Lake–Heaven Lake greenstone belt (Thurston et al. 1987; *see* Figure 9.2). Spinifex-textured komatiite flows and ultramafic pyroclastic rocks generally occur in diverse volcanic sequences with primary structures indicative of a shallow-water, platform-facies depositional environment (Thurston and Chivers 1990). The approximately 3 billion-year-old age of the Lumby Lake greenstone belt (Davis and Jackson 1988) suggests that all these sequences may be Mesoarchean in age.

DIVERSE INTERMEDIATE TO FELSIC VOLCANIC SEQUENCES

The diverse volcanic sequences within the Wabigoon Subprovince are recognized (Trowell et al. 1980; Johns and Thurston 1987) across the western and eastern regions (Table 9.5), but their stratigraphic settings are in contrast. In the WWR, the diverse volcanic sequence is composed predominantly of intermediate to felsic volcaniclastic rocks, pyroclastic rocks and flows with subordinate mafic clastic and pyroclastic rocks, mafic and ultramafic flows. The diverse volcanic sequences overlie the lower mafic sequences either conformably or unconformably. Diverse volcanic sequences comprise: 1) sequences representing complex stratigraphy with little discernable order, which are considered single complex units at the group level; 2) a basal complex unit with little discernable order accompanied by a more homogeneous unit consisting of a small range of rock types, commonly described as 2 groups or formations; and 3) a sequence with readily apparent stratigraphic order comprising 1 or more mafic to felsic metavolcanic cycles. Most of these diverse volcanic sequences are complexly folded so that stratigraphic thicknesses cannot in all cases be deduced,

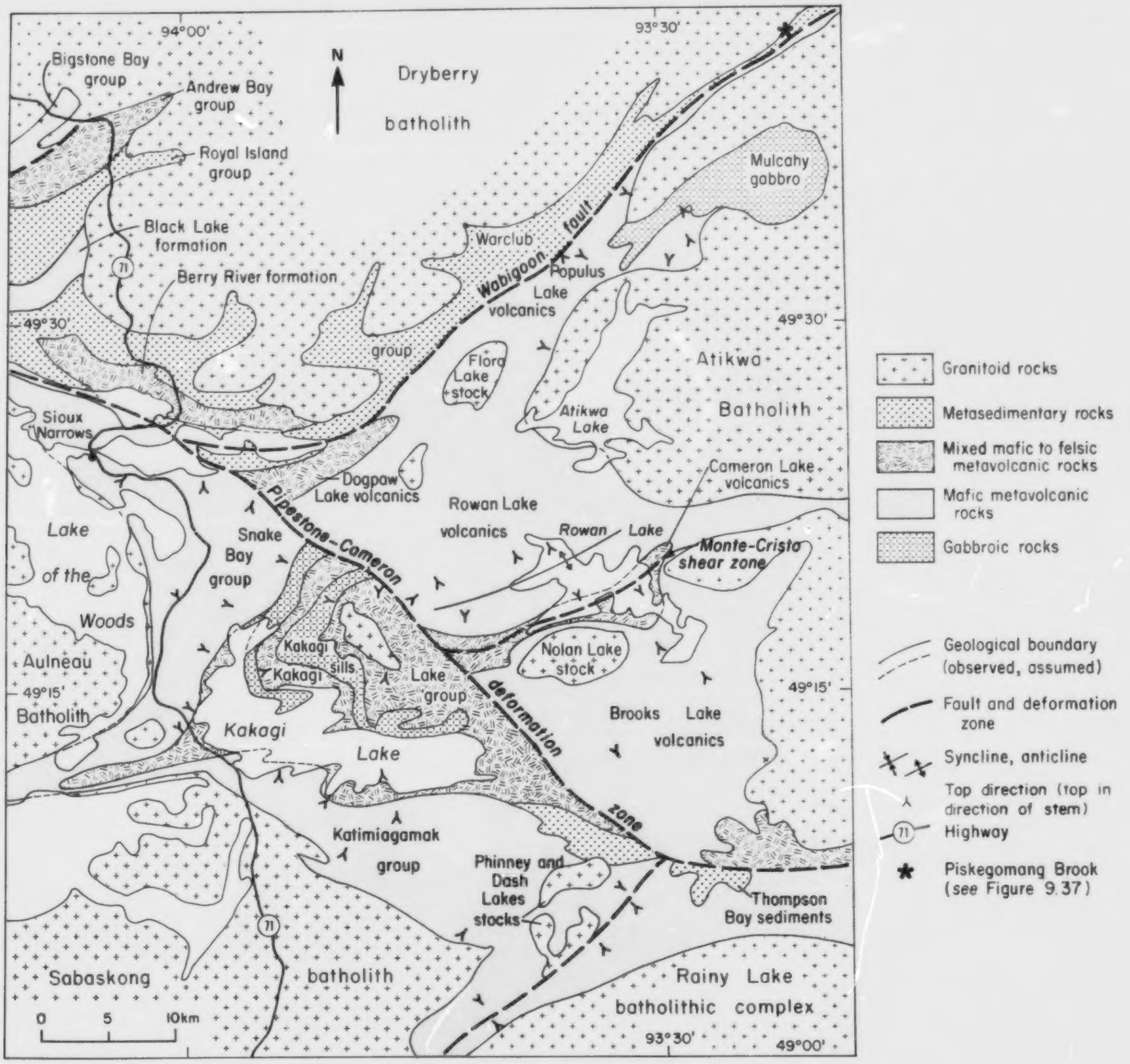

94°00'
93°30'
N
Dryberry
batholith
Bigstone Bay group
Andrew Bay group
Royal Island group
Black Lake formation
71
Berry River formation
Warclub group
Wabigoon fault
Populus Lake volcanics
Mulcahy gabbro
49°30'
Flora Lake stock
Atikwa Lake
Atikwa Batholith
Sioux Narrows
Dogpaw Lake volcanics
Rowan Lake volcanics
Rowan Lake
Cameron Lake volcanics
Monte-Cristo shear zone
Pipestone–Cameron deformation zone
Lake of the Woods
Snake Bay group
Kakagi sills
Kakagi Lake group
Nolan Lake stock
Aulneau Batholith
49°15'
Kakagi Lake
Brooks Lake volcanics
Katimiagamak group
Phinney and Dash Lakes stocks
Thompson Bay sediments
71
Sabaskong batholith
Rainy Lake batholithic complex
0
5
10km
93°30'
49°00'
Granitoid rocks
Metasedimentary rocks
Mixed mafic to felsic metavolcanic rocks
Mafic metavolcanic rocks
Gabbroic rocks
Geological boundary (observed, assumed)
Fault and deformation zone
Syncline, anticline
Top direction (top in direction of stem)
Highway
Piskegomang Brook (see Figure 9.37)

Figure 9.10. Detailed geology, Kakagi–Rowan Lakes greenstone belt.

but they are consistently less thick than the lower mafic volcanic rocks. For example, the top 3 formations of the Kakagi Lake group (*see* Figure 9.10; Emm Bay, Cedartree Lake and Stephen Lake formations) form a homoclinal succession of diverse volcanic sequence volcanic rocks 5 to 6 km in thickness.

Though not laterally continuous throughout the WWR, individual sequences occupying similar stratigraphic positions have been identified (*see* Figures 9.2, 9.8, 9.9, 9.10, 9.12 and 9.13). Whereas these individual sequences are similar in rock type, chemistry, depositional environment and mechanisms of eruption and deposition, they are not contemporaneous.

The most common volcanic deposits are primary and resedimented pyroclastic rocks, clastics, flows and domes of intermediate to felsic composition, with smaller amounts of mafic and mafic to intermediate flows and felsic pyroclastic rocks and flows. Resedimented pyroclastic rocks, for example those in the Kakagi Lake group (Johns 1985; Edwards 1980; Kaye 1974a, 1974b), comprise poorly sorted heterolithic debris flows. Lithic fragments within the matrix are commonly similar to the larger clasts and both are derived from all rock types making up the diverse volcanic sequence volcanic rocks. The debris flow deposits are thick, massive beds that may exhibit both reverse and normal grading.

Primary intermediate pyroclastic deposits are well to poorly sorted, monolithic, and are composed of subangular to angular clasts supported in a matrix similar in composition to that of the clasts. Depositional mechanisms include:

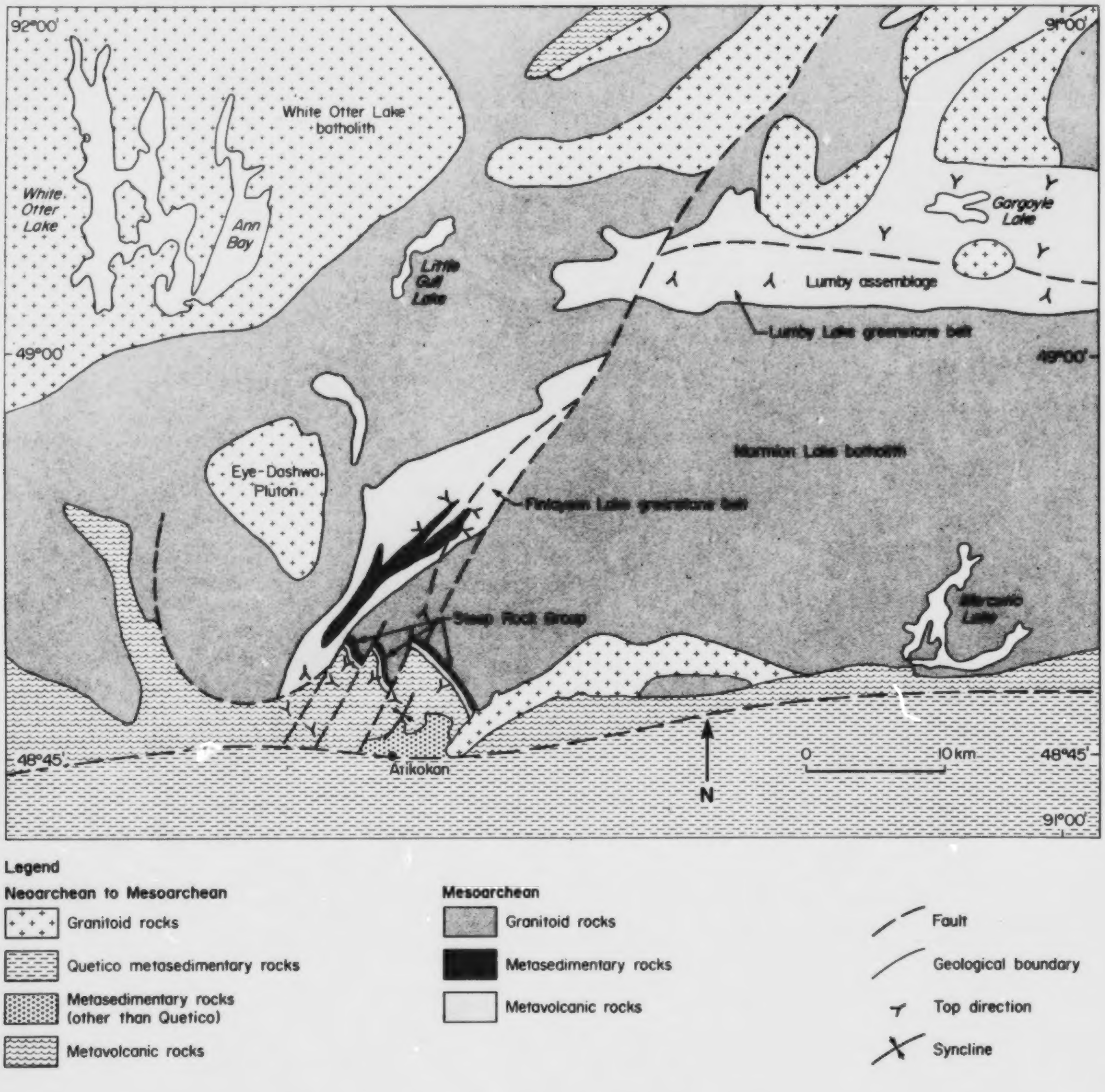

92°00'
91°00'
White Otter Lake batholith
White Otter Lake
Ann Bay
Little Gull Lake
Gargoyle Lake
Lumby assemblage
Lumby Lake greenstone belt
49°00'
49°00'
Marmion Lake batholith
Eye-Dashwa Pluton
Finlayson Lake greenstone belt
Steep Rock Group
48°45'
48°45'
Atikokan
0
10 km
N
91°00'
Legend
Neoarchean to Mesoarchean
Granitoid rocks
Quetico metasedimentary rocks
Metasedimentary rocks (other than Quetico)
Metavolcanic rocks
Mesoarchean
Granitoid rocks
Metasedimentary rocks
Metavolcanic rocks
Fault
Geological boundary
Top direction
Syncline

Figure 9.11. Detailed geology, Lumby–Finlayson Lakes greenstone belt.

subaqueous pyroclastic flow, (e.g., Berry River formation [Johns 1987]; Figure 9.20), fall-out, base surge, phreatic and phreatomagmatic explosions, (Figure 9.21; e.g., Monument Bay formation [Morrice 1989]) and autoclastic brecciation (e.g., Emm Bay formation [Johns 1985]). Distinctive euhedral plagioclase-phyric flows and primary pyroclastic deposits form homogeneous assemblages in the Rowan Lake (Johns 1986) and Yoke–Straw lakes areas (Edwards 1983b) and occur interdigitated with other rocks elsewhere. Common features in primary deposits are the presence of pumice (Figures 9.22 and 9.23), normal and reverse grading, double grading, cross-bedding, hydrothermal alteration, gas-escape structures (*see* Figure 9.20), bomb sags, and fiamme (cf., Johns 1985).

Mafic to intermediate flows are both aphyric and porphyritic, with euhedral plagioclase laths. Large amygdules and gas cavities are common in massive flows, pillowed flows, flow breccia, pillow breccia and hyaloclastite (Figure 9.24). Examples include northern Lake of the Woods (Ayer 1987), the Neepawa group (Berger 1988a), the Manitou Straits–Straw Lake area (Berger 1988b; Edwards 1983b), the Manitou Lake area (Blackburn 1982) and the Wabigoon Lake area (Satterly 1943). Pillows can measure up to 10 by 4m with thick selvages of hyaloclastite. In the Neepawa group, some massive and pillowed mafic flows contain varioles derived by liquid immiscibility (cf., Gelinas et al. 1976) that coalesce at the tops of flows (Berger 1988a).

Quartz- and feldspar-phyric felsic lavas are a lesser component of the diverse volcanic sequence volcanic rocks. Flow and dome complexes include flow-banded units, some of which contain autoclastic breccia and pyroclastic rocks (*see* Figure 9.23; cf., Easton and Johns 1985). Examples include the Off Lake–Burditt Lake (Blackburn 1976b) and the Phinney–Dash lakes (Edwards 1983a) areas. Pyroclastic units and lava flows occur as subaqueous and subaerial deposits. Welding and deposition by the ash-flow mechanism in a shallow-water to subaerial environment is indicated by the presence of: pumice and devitrification spherulites, (e.g., Stevens Lake formation [Johns 1985]), and eutaxitic textures and flow banding as in the Upper Member of the Cameron Lake volcanics (Johns 1986), and in the Rat Portage Bay area, on Lake of the Woods (Ayer 1987).

Intravolcanic metasedimentary rocks are found within both mafic to felsic volcaniclastic and pyroclastic sequences and mafic flows such as the upper part of the Snake Bay group. The intravolcanic metasedimentary rocks are generally relatively thin, quartz-poor, feldspathic wacke and reworked tuff interbedded with siltstone, chert, siliceous tuffs as well as oxide and sulphide iron formation. In the Kakagi Lake group, thicker units of fine cherty siltstone and chert are interbedded with fine feldspathic wacke to arenite and well-bedded pyroclastic rocks (Davies and Morin 1976; Kaye 1974a, 1974b). Some cherty tuff and siliceous siltstone beds form gradational caps to felsic ash-flows

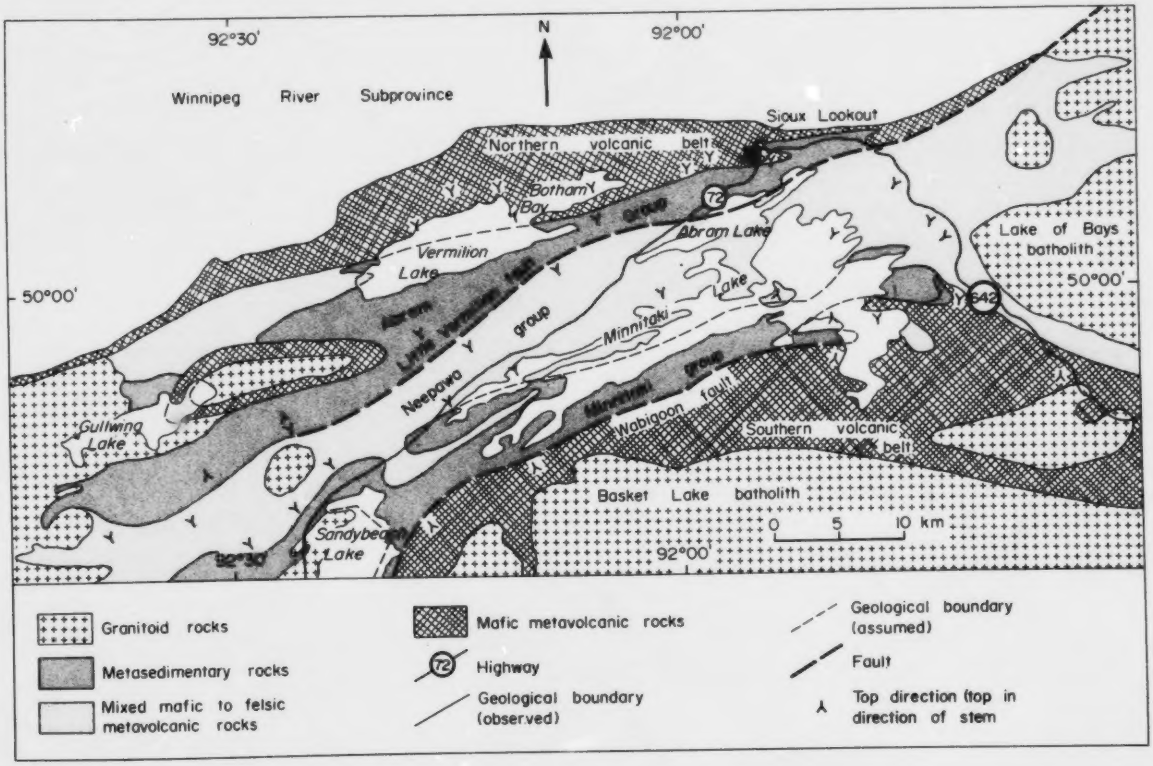

92°30'
N
92°00'
Winnipeg River Subprovince
Sioux Lookout
Northern volcanic belt
Botham Bay
72
Abram Lake
group
Vermilion Lake
Abram Lake–Vermilion fault
Lake of Bays batholith
50°00'
50°00'
642
Minnitaki Lake
Neepawa group
Minnitaki group
Wabigoon fault
Gullwing Lake
Southern volcanic belt
Basket Lake batholith
0 5 10 km
92°00'
Sandybeach Lake
92°30'
Granitoid rocks
Metasedimentary rocks
Mixed mafic to felsic metavolcanic rocks
Mafic metavolcanic rocks
Highway
Geological boundary (observed)
Geological boundary (assumed)
Fault
Top direction (top in direction of stem)

Figure 9.12. Detailed geology, Abram Lake greenstone belt and Minnitaki Lake greenstone belt.

indicating that many of these epiclastic units are the lateral equivalents of felsic volcanism (*see* Figure 9.23). Thick cherts within bedded metasedimentary rocks (Phinney–Dash lakes volcanic centre [Edwards 1980, 1983a]) indicate a period of quiescence. In the Dogpaw Lake area (Johns 1984), a thin volcaniclastic conglomerate is interbedded with primary intermediate pyroclastic deposits and flows indicating fluviatile reworking.

Distinctive clinopyroxene (now hornblende)-phyric flows and pyroclastic rocks occur within the diverse volcanic sequences in the Lake of the Woods area (*see* Figures 9.9 and 9.25; Morrice 1988, 1989; Ayer 1988b; Ayer and Buck 1989; Trowell 1986b; Johns 1987) and in the Sioux Lookout area (Berger 1987, 1988b; Chorlton 1991). Intermediate and mafic rocks of the unit are vesicular, with pahoehoe toes, bomb sags, and base-surge cross-bedding indicating deposition by shallow-water to subaerial phreatic eruptions, possibly in a cinder cone environment (*see* Figure 9.21; Morrice 1989).

The rocks of the diverse volcanic sequences represent subaqueous and subaerial environments. The underlying mafic metavolcanic sequences form shallowing-upward subaqueous units; evidence of continued deposition in shallow water is found throughout the diverse volcanic sequences volcanic assemblages as described in the following paragraphs.

The shallow-water to subaerial nature of the contact between the mafic Snake Bay group and the overlying diverse volcanic sequences Emm Bay formation has already been described here (Johns 1985). Structures, textures, and rock types indicative of shallow-water deposition commonly found within the diverse volcanic sequences volcanic rocks are: abundant amygdules and gas cavities, pillow breccia-hyaloclastite, fluviatile metasedimentary rocks, base-surge cross-bedding, mafic scoria and felsic pumice. The following criteria indicate subaerial deposition of some of the felsic rocks of the diverse volcanic sequences: devitrification spherulites, eutaxitic textures and gas-escape structures in felsic tuffs, ground-surge and ash-cloud deposits in block and ash flows, pahoehoe toes in mafic flows, highly amygdaloidal massive mafic and intermediate flows, and bomb-sags in mafic tuffs comprising cinder cones. The Stephens Lake formation of the Kakagi Lake group and the Monument Bay group, Lake of the Woods (Morrice 1989), contain many of these features.

Volcaniclastic rocks of the diverse volcanic sequences represent resedimented and primary volcanic units. Following the initial study of Johns (1985) recent mapping has led

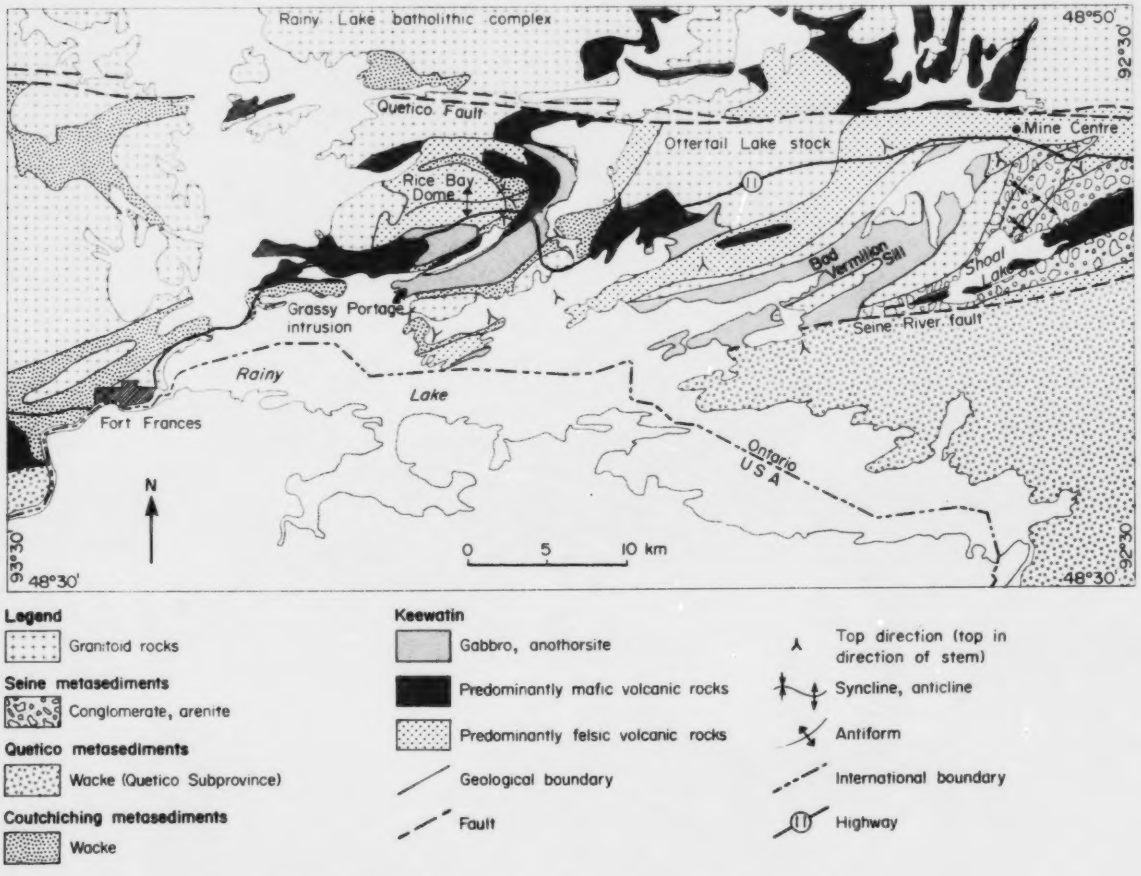

Rainy Lake batholithic complex
48°50'
92°30'
Quetico Fault
Ottertail Lake stock
Mine Centre
Rice Bay Dome
11
Bad Vermilion Sill
Shoal Lake
Seine River fault
Grassy Portage intrusion
Rainy
Lake
Fort Frances
Ontario
U S A
N
0 5 10 km
93°30'
48°30'
92°30'
48°30'
Legend
Granitoid rocks
Seine metasediments
Conglomerate, arenite
Quetico metasediments
Wacke (Quetico Subprovince)
Coutchiching metasediments
Wacke
Keewatin
Gabbro, anothorsite
Predominantly mafic volcanic rocks
Predominantly felsic volcanic rocks
Geological boundary
Fault
Top direction (top in direction of stem)
Syncline, anticline
Antiform
International boundary
Highway

Figure 9.13. Detailed geology, Fort Frances–Mine Centre greenstone belt.

to an interpretation of the genesis of many volcaniclastic units (e.g., Ayer et al. 1987; Ayer and Buck 1989). The resedimented rocks are relatively coarse (Figure 9.26), massive units commonly displaying normal or reverse grading, and are overlain vertically or laterally by finer units, mainly wackes with Bouma divisions. They are interpreted as turbidity current deposits.

Primary volcaniclastic rocks represent several deposit types. Doubly graded, monolithic to bilithic, moderately to poorly sorted, coarse deposits were deposited by subaqueous pyroclastic flows (e.g., Johns 1985). Monolithic, well- to moderately sorted, coarse to fine deposits that exhibit a normal grading, reverse grading of pumice, ground surge and ash-top deposits, eutaxitic

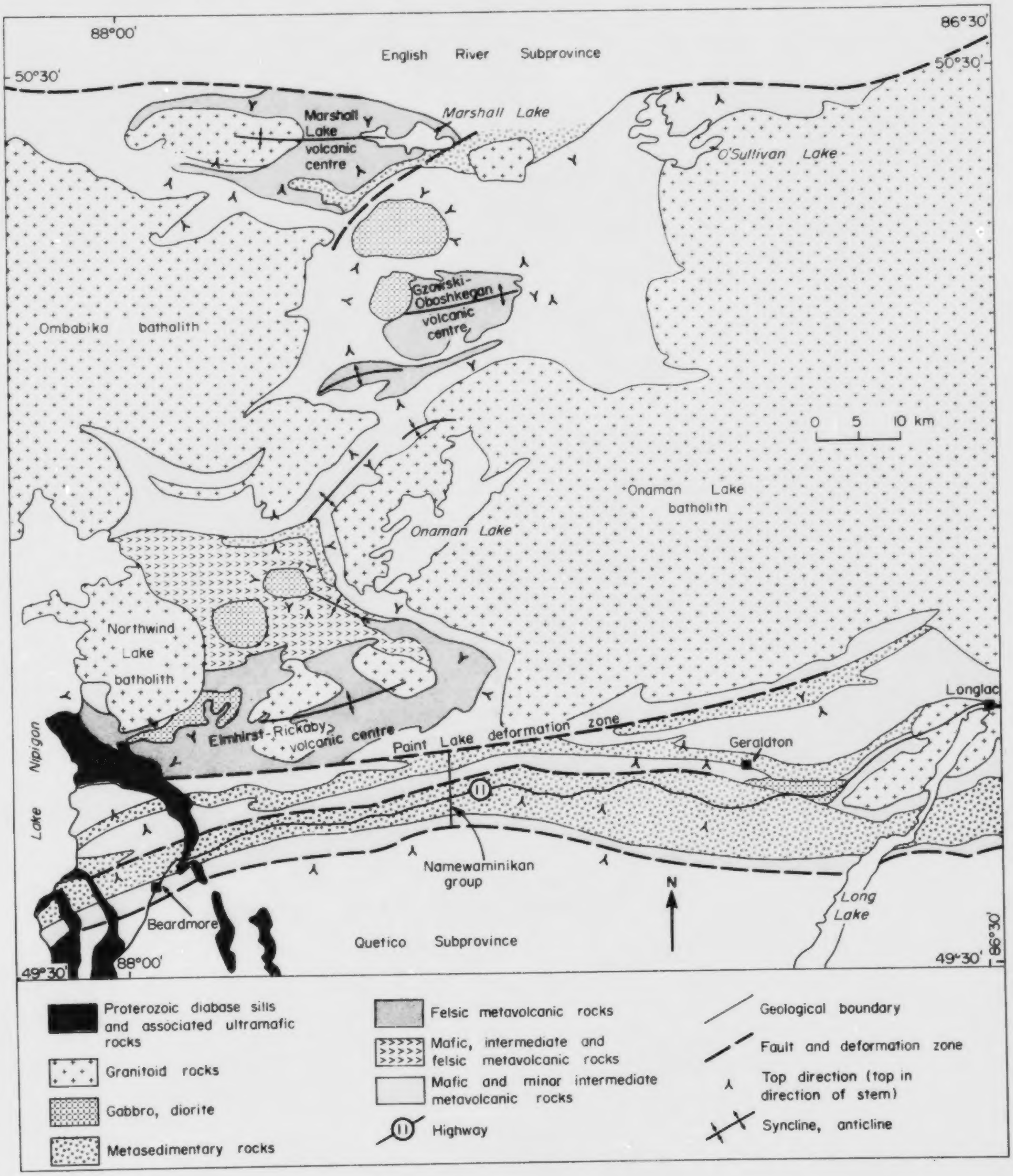

88°00'
86°30'
50°30'
50°30'
English River Subprovince
Marshall Lake volcanic centre
Marshall Lake
O'Sullivan Lake
Gzowski-Oboshkegan volcanic centre
Ombabika batholith
0 5 10 km
Onaman Lake batholith
Onaman Lake
Northwind Lake batholith
Elmhirst-Rickaby volcanic centre
Paint Lake deformation zone
Longlac
Geraldton
11
Lake Nipigon
Namewaminikan group
Beardmore
N
Long Lake
Quetico Subprovince
49°30'
88°00'
49°30'
86°30'
Proterozoic diabase sills and associated ultramafic rocks
Granitoid rocks
Gabbro, diorite
Metasedimentary rocks
Felsic metavolcanic rocks
Mafic, intermediate and felsic metavolcanic rocks
Mafic and minor intermediate metavolcanic rocks
Highway
Geological boundary
Fault and deformation zone
Top direction (top in direction of stem)
Syncline, anticline

Figure 9.14. Detailed geology, Beardmore–Geraldton Belt.

Table 9.3. Lower mafic metavolcanic sequences, western Wabigoon: composition, rock types, textures and superpositionary relationships.

Sequence (text figures)	Units	Depositional Environment	Rock Types and/or Primary Structures	Geochemistry	Boundary Relationships	Major References	U-Pb Age (Ma) Geochronology References
Deception Bay group (Fig. 9.9)		deep-water submarine	massive, pillowed, aphyric flows; minor plagioclase-megaphyric flows	mixes Mg and Fe tholeiites	lower contact tectonic with Winnipeg River Subprovince; upper contact conformable into Clearwater Bay group	Ayer 1990	
Bigstone Bay group (Fig. 9.9)	4 unnamed formations (A-base to D-top)	deep-water grading to shallow submarine near top of group	massive, pillowed, aphyric flows; extensive horizons of plagioclase-megaphyric flows; differentiated sills and/or flows	Fm. D—mixed Fe and Mg tholeiites Fm. C—Fe tholeiites Fm. B—mixed Fe and Mg tholeiites Fm. A—Mg tholeiites	lower contact intrusive with Dryberry Batholith; upper contact conformable into Andrew Bay group in the south and Clearwater Bay group in the north	Ayer 1990	
Cedar Island group (Fig. 9.9)		deep-water submarine	massive, pillowed, aphyric flows; extensive plagioclase-megaphyric flows; differentiated sills; minor felsic volcanic rocks	mixed Mg and Fe tholeiites, with komatiitic basalts	lower contact unexposed; upper contact conformable into Andrew Bay group	Smith 1987 Ayer 1990	
Barrier Islands group (Fig. 9.9)		deep-water submarine	massive, pillowed aphyric flows; extensive plagioclase-megaphyric flows; mafic sills; minor felsic volcanic rocks in the western part	mostly Fe tholeiites; minor Mg tholeiites	lower contact unexposed; upper contact conformable into Andrew Bay group to the north and the Monument Bay group to the south	Ayer 1990	
Snake Bay group (Figs. 9.9 and 9.10)	2 unnamed formations (upper and lower)	deep-water submarine	massive, pillowed, aphyric flows; extensive plagioclase-megaphyric flows in upper and lower formations; mafic sills	upper formation mostly Fe tholeiites; lower formation mostly Mg tholeiites	lower contact intrusive with the Aulneau Batholith; upper contact conformable into Long Bay group in west, unconformable into Kakagi Lake group in east	Morrice 1977 Johns 1987 Ayer 1990 Blackburn et al. 1985	
Katimiagamak group (Fig. 9.10)		submarine shield volcano	pillowed and massive, amygdaloidal aphyric flows; minor plagioclase-megaphyric flows; gabbro sills	Fe tholeiites	lower contact intrusive with the Sabaskong batholith; disconformably overlain by Kakagi Lake group	Edwards 1980 Trowell et al. 1980 Blackburn et al. 1985	$2731.7^{+4.0}_{-2.9}$ Davis and Edwards 1986
Rowan Lake volcanics–Populus Lake volcanics, Brooks Lake volcanics (Fig. 9.10)		shallow submarine near the top of the group	pillowed to massive aphyric flows; minor amygdaloidal and plagioclase megaphyric flows; intercalated intermediate pyroclastic rocks in upper part	predominantly tholeiites but intermixed with calc-alkalic volcanic rocks in the upper part	lower contact intrusive with the Atikwa–Lawrence batholith; upper contact conformable with the Cameron Lake volcanics	Davies and Wotowich 1958 Trowell et al. 1980 Johns 1985	

Table 9.3. Lower mafic metavolcanic sequences, western Wabigoon: composition, rock types, textures and superpositionary relationships.

Sequence (text figures)	Units	Depositional Environment	Rock Types and/or Primary Structures	Geochemistry	Boundary Relationships	Major References	U-Pb Age (Ma) Geochronology References
Eagle Lake group (Fig. 9.8)		submarine	massive and pillowed flows	mixed Fe and Mg tholeiites	lower contact intrusive with the Atikwa Batholith; unconformably overlain by the Lower Wabigoon volcanics	Moorhouse 1941 Blackburn et al. 1985	
Blanchard Lake group (Fig. 9.8)		submarine basaltic platform	massive and pillowed aphyric flows; minor plagioclase-phyric flows	predominantly Fe tholeiites in lower part and Mg tholeiites in upper part	lower contact not exposed; conformably overlain by the Upper Manitou Lake group	Blackburn 1982 Blackburn et al. 1985	
Wapageisi group (Fig. 9.8)	Starshine Lake subgroup—7 numbered formations	submarine basaltic platform	massive and pillowed aphyric flows with extensive formations of plagioclase-megaphyric flows	lower portion Mg-rich tholeiites and minor komatiitic basalts; upper part Fe-rich tholeiites	lower contact intrusive with the Meggisi pluton; upper contact conformable into the Etta Lake formation	Blackburn 1982	
	Etta Lake formation		massive aphyric and plagioclase-phyric mafic flows interbedded with clastic sedimentary and pyroclastic volcanic rocks		lower contact conformably overlying Starshine Lake subgroup; upper contact conformable into the Glass Bay formation	Blackburn et al. 1985	
	Glass Bay formation		massive, pillowed, pillowed brecciated aphyric flows		lower contact conformably overlying Glass Bay formation; upper contact conformable into the Manitou Lakes group		
Botham Bay group (Fig. 9.12)		submarine	pillowed and massive flows with interbedded magnetite iron formation near the top of the unit	mixed Mg tholeiites, Fe tholeiites and minor komatiites	lower contact intruded by Winnipeg River Subprovince granitoids; unconformably overlain by the Patara group	Johnston 1969, 1972 Trowell et al. 1978 Trowell et al. 1980	
South Sturgeon Lake volcanics (Fig. 9.3)	Darkwater Lake cycle—formation A (lowest of 3 unnamed formations in the cycle)	broad submarine shield volcano	pillowed and massive; aphyric and plagioclase-phyric mafic flows; locally interbedded clastic and pyroclastic units	mixed tholeiitic and calc-alkalic rocks	lower contact intruded by the Southern Granitic Complex; upper contact transitional into pyroclastic rocks and debris flows of formation B—Darkwater Lake cycle	Trowell 1983	2736 ±2 Davis et al. 1985
Jutten group (Fig. 9.3)		submarine	pillowed and massive, aphyric, mafic flows, minor porphyritic mafic flows and flow breccia; interbedded magnetite iron formation unit near base of the group in the west and conglomerated	mixed Mg and Fe tholeiites and komatiitic basalts	lower contact intruded by various granitoids; upper contact unconformably overlain by the Savant Narrows formation	Trowell 1986a	

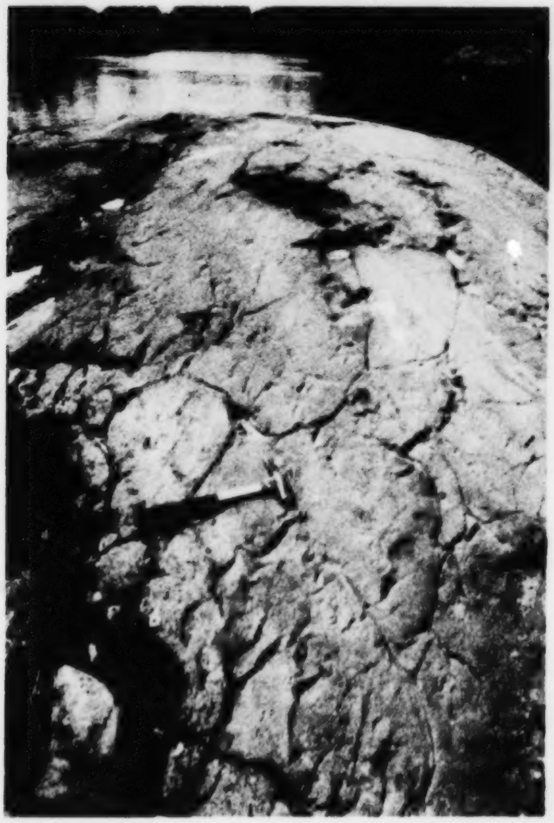


Figure 9.15. Pillowed ultramafic flow from the Cliff Island formation, Coste Island, Lake of the Woods. Pillows with very thin selvages and orange weathering are typical of the upper mafic to ultramafic volcanic sequences in Lake of the Woods (photograph *courtesy of* J. Ayer).

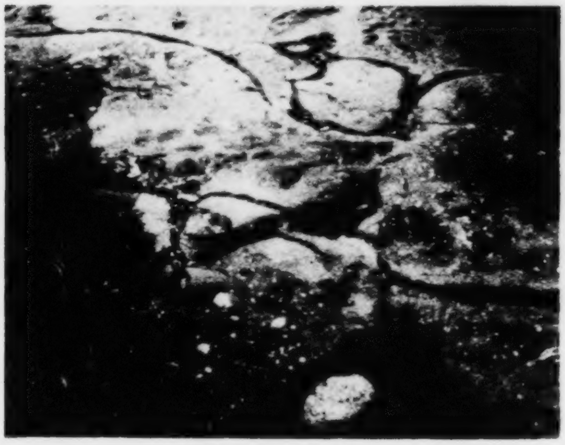


Figure 9.17. Pillowed plagioclase megaphyric basalt flows in the Barrier Islands group, East Allie Island, Lake of the Woods (photograph *courtesy of* J. Ayer).

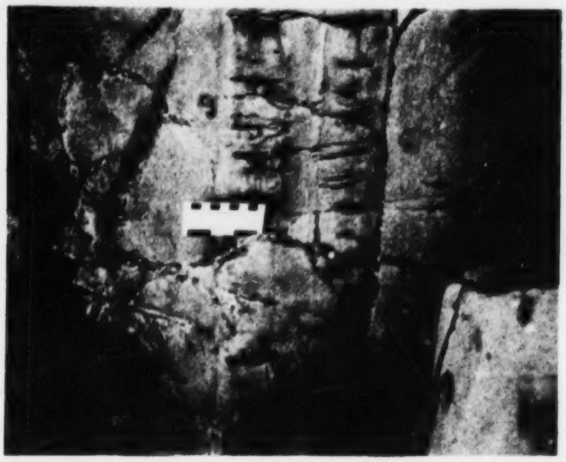


Figure 9.18. Mafic feeder dike intruding mafic flows of the Snake Bay group near Sioux Narrows (photograph *courtesy of* G. Johns).

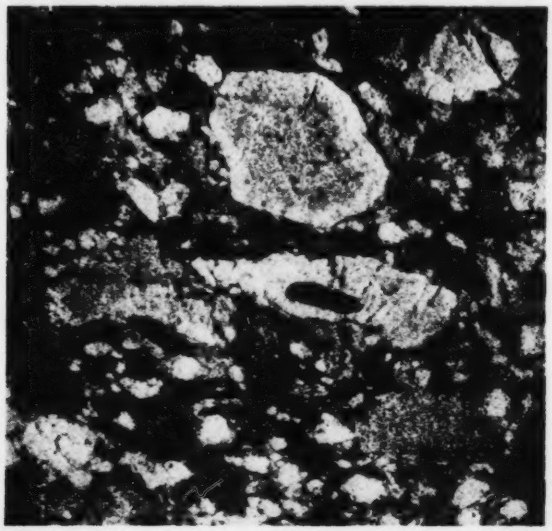


Figure 9.16. Calc-alkalic andesite isolated to broken pillow breccia, Lower Manitou Lake area (photograph *courtesy of* G. Johns).

textures and gas-escape structures, are interpreted as ash-flow deposits. Well-bedded, fine-grained beds with diffuse contacts are interpreted to have been deposited by fall-out from tephra clouds (cf., Fisher and Schmincke 1984, p.126). Trough cross-bedded mafic units are the product of base-surges and phreatomagmatic eruptions (cf., Fisher and Schmincke 1984).

Rare earth element (REE) data for both calc-alkalic and tholeiitic rocks in the diverse volcanic sequences show an overall enrichment of up to 100 times chondrites in the light rare earth elements (LREE) over rocks in the lower mafic group. The felsic rocks are more strongly LREE enriched and heavy rare earth element (HREE) depleted than the calc-alkalic andesitic to basaltic flows. A notable exception is the South Sturgeon Lake volcanics (Lesher et al. 1986). Felsic volcanic rocks in the Neepawa volcanics have HREE abundances 2 to 4 times chondrites (Berger 1988a). The

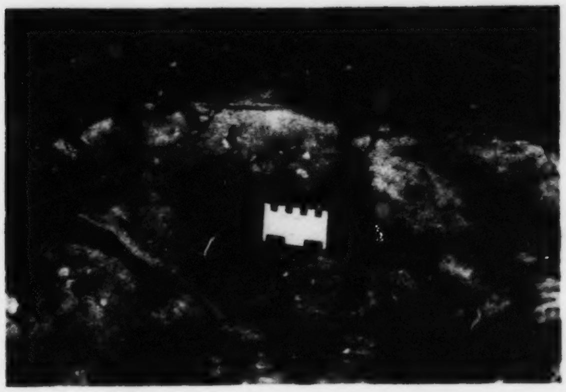


Figure 9.19. Exposed base of a differentiated sill in contact with pillowed plagioclase megaphyric basalt flow in the Bigstone Bay group, Hades Islands, Lake of the Woods. Note the chilled base of the sill follows the contours of the pillow rather than a sharp crosscutting contact. The sill grades upwards into medium-grained, locally layered, peridotite (photograph *courtesy of* J. Ayer).

tholeiitic rocks have a lesser enrichment compared to the more intermediate calc-alkalic rocks. In general, the REE data indicate that felsic metavolcanic rocks cannot be derived from the mafic units by reasonable degrees of fractionation of the observed phenocryst phases, in common with Archean metavolcanic units elsewhere in the Canadian Shield (cf., Thurston and Fryer 1983).

Mafic dikes intruding the intermediate volcanic rocks have REE patterns similar to the mafic flows and are therefore considered feeders to them. Felsic ash flows (e.g., Ayer 1988a) have a HREE enrichment and strong negative europium anomalies and may have been erupted from a zoned magma chamber (cf., Hildreth 1979).

The bulk of the high-precision radiometric ages in the Wabigoon Subprovince have been obtained on units within the diverse volcanic sequences. This geochronologic evidence, as well as stratigraphic position, indicate that many of the diverse volcanic sequences are not synchronous, but may have been deposited in geographically and temporally distinct volcanic centres. U-Pb dating of zircon indicates that diverse volcanic sequences in the WWR range from 2745 Ma for the Handy Lake group (Davis and Trowell 1982) to 2711 Ma for a unit near the top of the Kakagi Lake group. However, most of this group is intruded by 2725 million-year-old mafic sills (Davis and Edwards 1982). Further details are provided in a later section of this chapter (*see* Discussion *in* Metasedimentary Rocks).

Metasedimentary Rocks

Metasedimentary rocks of the Wabigoon Subprovince largely occur as thick, laterally extensive units with minor, local metavolcanic lenses. The major sequences are listed in Table 9.6. Clastic metasedimentary rocks dominate, with a full range of grain sizes observed. Chemical metasedimentary rocks comprise iron formation, with smaller volumes of chert-forming interflow units. An unique clastic-chemical sedimentary sequence (Steep Rock Group) contains basal clastics overlain by stromatolite-bearing carbonates, and various oxide iron formations.

Clastic sedimentary sequences in the Superior Province have classically been grouped into 2 types:

1. "Keewatin"-type sequences (Lawson 1885) form an integral part of the volcanic succession lying conformably on, or interdigitating with, volcanic rocks, and showing a similar transition at their upper boundaries. Wackes and mudstones predominate, coarser units being subordinate (e.g., Thurston and Chivers 1990)
2. "Timiskaming"-type units (cf., Thomson 1946; Hewitt 1963; Cooke and Moorehouse 1969) lie unconformably on "Keewatin" volcanic rocks. Conglomerates predominate over finer units and, in addition, alkaline volcanism is commonly closely associated with these sequences (e.g., Cooke and Moorhouse 1969; Shegelski 1980; Blackburn 1982). "Timiskaming"-type sequences are considered to be largely fluviatile in origin (Hyde 1980; Shegelski 1980; Blackburn 1982) and are marked (Thurston and Chivers 1990) by rapid lateral and vertical facies changes. Hence, both alluvial fan-fluvial and resedimented facies associations are found in juxtaposition (Shegelski 1980; Hyde 1980; Krapez and Barley 1987).

CLASTIC SEDIMENTARY SEQUENCES

The sequences may be grouped into west-southwest-trending lithologic associations, commencing at the west end of the western Wabigoon domain: the Crowduck Lake group and White Partridge Bay group, in northern Lake of the Woods (*see* Figure 9.9); the Warclub–Minnitaki metasediments and Abram group, from Lake of the Woods to Sioux Lookout (*see* Figures 9.8, 9.9, 9.10 and 9.12); the Savant Lake group, again near the northern boundary (*see* Figure 9.3); the Thompson Bay–Esox Lake–Manitou Lakes group–Stormy Lake groups, through the central part of western Wabigoon (*see* Figure 9.10); the Quest Lake metasediments, at Sturgeon Lake (*see* Figure 9.3); the Coutchiching and Seine groups, along the southern boundary of the subprovince (*see* Figure 9.13); and the sedimentary units of the Namewaminikan group, (Devaney and Williams 1989) in the Beardmore–Geraldton Belt (*see* Figure 9.14).

CHEMICAL METASEDIMENTARY ROCKS

Chemical metasedimentary rocks of the subprovince include ubiquitous iron formation and rare, scattered occurrences of marble. Iron formation could represent a variety of depositional environments: submarine fan (Meyn and Palonen 1980), shallow water platforms (Blake 1983; Morris and Horwitz 1983) and deep marine (Shegelski 1978; Hoffman 1987). Magnetite-bearing iron formation is closely associated with wackes of some of the clastic sedimentary sequences . Iron formation laminae are interbedded with turbidites in the Minnitaki (*see* Figure 9.12), Manitou, Stormy Lake, Namewaminikan and Savant Lake groups and in some cases are an integral part of Bouma-type graded beds; the magnetite layers occur between graded wacke

Table 9.4. Upper mafic metavolcanic sequences, western Wabigoon: composition, rock types, textures and superpositionary relationships.

Sequence (text figures)	Units	Depositional Environment	Rock Types and/or Primary Structures	Geochemistry	Boundary Relationships	Major References	U–Pb Age (Ma) Geochronology References
Clearwater Bay group (Fig. 9.9)	High Lake formation	submarine	massive and pillowed aphyric mafic flows		conformably underlain and overlain by other formations of the Clearwater Bay group	Ayer 1990	
Indian Bay group (Fig. 9.9)		shallow submarine	massive and pillowed aphyric mafic flows, minor felsic to intermediate pyroclastic rocks	Fe-tholeiitic and minor calc-alkalic volcanic rocks	conformably overlain by Clearwater Bay group in north and Andrew Bay group in the south	Smith 1987 Ayer 1988b Ayer 1990	
Monument Bay group (Fig. 9.9)	Cochrane Island formation	shallow submarine	pillowed and massive aphyric mafic and ultramafic flows	Mg-tholeiitic and komatiitic basalt	conformably underlain and overlain by predominantly calc-alkalic formations of the Monument Bay group	Morrice 1989 Ayer and Buck 1989	
	Black Lake formation	submarine	massive and pillowed, aphyric mafic volcanic rocks, minor intermediate pyroclastic rocks	Mg-tholeiitic and komatiitic basalt	conformably overlain by Warclub group sedimentary rocks	Johns 1987 Ayer and Buck 1989	
Long Bay group (Fig. 9.9)	Cliff Island formation	shallow submarine	pillowed and massive aphyric mafic and ultramafic flows with minor intercalated pyroclastic rocks and siltstones	Mg-tholeiitic basalt, komatiitic basalt and ultramafic komatiite	conformably underlain and overlain by calc-alkalic volcanic formations of the Long Bay group	Ayer, in press	
Upper Wabigoon volcanics (Fig. 9.8)		submarine	pillowed flows with minor intercalated tuffs and cherty interflow sedimentary rocks	predominantly Fe-tholeiitic, minor calc-alkalic rocks Mg-tholeiitic	conformably overlies the predominantly calc-alkalic volcanic rocks of the Lower Wabigoon volcanics; upper contact truncated by the Wabigoon fault	Blackburn et al. 1982	2743 ±1 2735^{+4}_{-3} Davis et al. 1982
Boyer Lake group (Fig. 9.8)		submarine	pillowed mafic flows with minor variolitic and plagioclase-phyric flows, minor felsic flows toward the top of the group	predominantly Fe-tholeiitic rocks with minor Mg-tholeiitic and calc-alkalic volcanic rocks	fault-bounded upper and lower contacts	Trowell et al. 1980 Blackburn 1982 Blackburn et al. 1982	2719 ±3 2722 ±5 Davis 1990

beds, or constitute the upper pelitic interval of such beds. Shegelski (1978) suggested a volcanogenic origin for the iron laid down as background sedimentation in turbidity-current fed basins. Iron formation clasts in resedimented conglomerates indicate active erosion of iron formation shortly after lithification.

The unusual Steep Rock Group (*see* Figure 9.11), near Atikokan, consists of stromatolite-bearing carbonate overlain by goethite- and hematite-rich breccia, the ore horizons of the Steeprock and Caland iron mines. Marble with possible stromatolites has been identified in association with iron formation in the Lumby Lake belt (Jackson 1985). Marbles are found in the Obonga Lake greenstone belt of the central Wabigoon domain (Kustra 1967).

METAVOLCANIC UNITS

Minor metavolcanic units occur within the metasedimentary units of the Wabigoon Subprovince. These units include alkalic metavolcanic rocks, plagioclase-phyric calc-alkalic units, and hornblende-phyric units. The major occurrence of alkalic metavolcanic rocks is the Sunshine Lake and Uphill Lake formations (Blackburn 1982) in the WWR. These formations are trachybasaltic and have associated metasedimentary and pyroclastic units representing a subaerial to shallow-water environment. Plagioclase-megaphyric andesite to rhyolite is associated with alluvial fan-fluvial metasedimentary rocks at Yoke Lake (Edwards 1983b) and with metasedimentary rocks at Esox Lake and Thompson Bay (Smith 1990).

Quartz-feldspar-phyric metavolcanic rocks are interbedded with the metasedimentary rocks of the Warclub group in the Lake of the Woods area (Johns 1987). The most extensive of these units is the Berry River formation which is conformably underlain and overlain by metawacke. The Berry River formation is a volcanic complex comprising central to distal depositional and redeposited volcanic facies.

FACIES ASSOCIATIONS

Two sedimentary facies associations (Walker 1979; Ojakangas 1985) predominate in the Wabigoon Subprovince metasedimentary units: resedimented or turbidite facies and alluvial fan-fluvial facies. Minor platform facies have been recognized in the EWR, in the Steep Rock Group (*see* Frontispiece *in* Wilks and Nisbet 1988) , and at Lumby Lake (Jackson 1985; Thurston and Chivers 1990), but this association has not been identified in the WWR.

The alluvial fan-fluvial facies association commonly underlies the turbidite association where they are in proximity. Examples include: 1) in the Sioux Lookout area, the alluvial Ament Bay formation is overlain by turbidites of the Daredevil and Little Vermilion formations (Turner and Walker 1973); and 2) in the Manitou Lake area, pyroclastic and alluvial rocks of the Uphill Lake formation are overlain by the Mosher Bay formation turbidites (Teal and Walker 1977; Teal 1979). This stratigraphic relationship reflects a subsiding basin (Teal and Walker 1977).

The lateral relationships between various facies associations in the western Wabigoon domain is problematical. Only in the Stormy Lake group can alluvial fan-fluvial facies (Figure 9.27) be traced laterally into turbidite facies (Kresz 1984). The alluvial fan-fluvial facies of the Seine metasediments has been interpreted (Wood 1980) to be the proximal equivalent of distal turbidites of the Quetico and Coutchiching sedimentary rocks. The Warclub–Minnitaki–Abrams sediments are dominantly turbidites at the east end (Figure 9.28; Walker and Pettijohn 1971), whereas towards the west end, near Sioux Narrows, they are intercalated with pyroclastic rocks of the Berry River formation and other unnamed units (Johns 1987). It has been suggested (Pettijohn 1935; Johnston 1969, 1972) that the Minnitaki group is the lateral equivalent of the Abram group. In this case, age data indicate that the intervening volcanic Neepawa group, which is about 30 million years older, has been tectonically positioned between the two (Davis et al. 1988).

Devaney and Fralick (1985) and Devaney and Williams (1989) have documented a complete transition from turbidites through a tidal facies to alluvial facies at the boundary between the EWR and the Quetico Subprovince. These metasedimentary rocks have been tectonically disrupted and are exposed as a series of thrust-bounded panels (Devaney and Williams 1989).

The facies associations of many of the WWR clastic sequences are not clear. The White Partridge Bay group has been considered resedimented (Thomson 1937) or alluvial fan-fluvial (Ayer 1987, 1988a; Figure 9.29). The Patara group is a complex mixture (N. Trowell, Ontario Geological Survey, personal communication, 1983) of volcanogenic clastic metasedimentary rocks, primary and redeposited volcanic units, and derived conglomerates. Other sedimentary sequences representing reworked pyroclastic units include the east end of the Kakagi Lake group (Edwards 1980), and the Handy Lake group metasedimentary rocks (Bond 1977, 1980). The Thompson Bay and Esox Lake sedimentary rocks (Edwards 1983b) are probably representative of the resedimented facies association. The Sturgeon–Quest–Post lakes sediments are insufficiently well known to determine facies association (Trowell 1976)

Paralic or shelf-facies metasedimentary rocks have been considered to be largely absent in greenstone belts (Ojakangas 1985). However, a shelf or platform environment has been suggested recently for the Steep Rock Group at Atikokan and the metasedimentary rocks of the Lumby Lake greenstone belt by Thurston and Chivers (1990). Geochronologic studies (Davis and Jackson 1988; D.W. Davis, Royal Ontario Museum, personal communication, 1990) indicate that both sequences are approximately 3 billion years old, or slightly younger in the case of the Steep Rock Group which lies unconformably over the 3 billion-year-old Marmion lake batholith.

Wilks and Nisbet (1988) described the Steep Rock Group as follows; a basal fining-upward clastic unit (Wagita Formation) which passes upward into laminated carbonates (Mosher Carbonate) containing diverse stromatolitic structures. The top of this formation is extremely irregular, suggesting (Jolliffe 1955) subaerial weathering and development of karst topography. The overlying goethite- and hematite-rich breccias and hematite pisolites of the Joliffe Ore Zone Formation may represent a paleosol of Archean (Riley 1970) or Mesozoic age (Machado 1987). The uppermost unit is an ultramafic pyroclastic rock (Dismal Ashrock Formation), capped by the Witch Bay Formation mafic to intermediate metavolcanic rocks. At Lumby Lake, iron formation, stromatolitic? marble, and quartz arenite lie above a mafic to felsic subaqueous volcanic sequence with rare spinifex-textured komatiite (Jackson 1985). However, the sequence at Lumby Lake may be intruded by the Marmion Lake batholith, whereas the Steep Rock Group unconformably overlies the same batholith, presenting as yet unresolved correlational problems.

Composition of Clastic Rocks

The provenance of sedimentary sequences of the Wabigoon Subprovince can be attributed to both volcanic and plutonic sources. Few detailed studies are available. Walker and Pettijohn (1971) infer a quartz porphyry source for wacke, and a granite terrain containing greenstones for arkose in the Sioux Lookout area. A number of workers have noted the paucity of both potassium feldspar grains in sandstones, and of potassium-feldspar-rich plutonic clasts in conglomerate. In contrast, sodium feldspar grains are abundant in sandstone, and tonalitic clasts have commonly been noted in conglomerate. Clast populations of conglomerate are commonly polymictic, with metavolcanic, metasedimentary and minor plutonic clasts.

MIGMATITES

Migmatitic metasedimentary rocks are absent from most of the Wabigoon Subprovince; this is a reflection of the low to medium metamorphic grade and the distribution of sedimentary rocks in the subprovince. Migmatitic sedimentary sequences are common in the Dryden area, marginal to the Winnipeg River Subprovince. Here, clastic metasedimentary rocks that are contiguous with the Warclub, Minnitaki and Abram groups display a northward progression from lower greenschist to upper amphibolite facies away from the Wabigoon fault, with isograds generally parallel to the fault (Bartlett 1983; Breaks et al. 1978; Breaks and Kuehner 1984; Breaks 1989). Metamorphic assemblages indicate an increase in pressure from less than 0.3 GPa, up to 0.4 GPa, in the temperature range 600° to 750°C (Bartlett 1978; Breaks 1989). The Ghost Lake Pluton is an S-type granite, derived from the enclosing metasedimentary rocks and is distinguished from other granitoids by its accessory mineral suite of sillimanite, garnet, tourmaline, cordierite, and rare beryl and dumortierite (a nesosilicate related to holtite) (Breaks et al. 1978; Breaks 1989).

DISCUSSION

The classification of metasedimentary sequences into Keewatin and Timiskaming has in the past been applied based mostly on these lithologic criteria. Examples of Timiskaming-type sequences within the WWR have been identified, largely on lithologic criteria, as follows: the Manitou series (Thomson 1934); the Seine metasediments (Davis et al. 1989); the White Partridge Bay group (Thomson 1937); and the Crowduck Lake group (Davies 1965). Structural studies (Hamilton and Hodgson 1984) have indicated the type Timiskaming to have been deposited after an early deformational event, but prior to, and thus affected by, a subsequent event. Using this more comprehensive definition, we consider the following metasedimentary sequences to be of the Timiskaming type: Crowduck Lake group (Figures 9.30 and 9.31; Davies 1965; Smith 1987; Ayer 1987), Esox Lake–Thompson Bay sediments (Edwards 1983b; Smith 1990; Edwards and Stauffer 1988), and the Seine metasediments (Davis et al. 1989). Sequences which have in the past been considered Timiskaming type, but which are of uncertain affiliation due to lack of structural studies are: Manitou Lakes group (Blackburn 1982), Minnitaki group (Walker and Pettijohn 1971), and the Stormy Lake group (McMaster 1978; Kresz 1984).

The various types of relationships between metasedimentary sequences, other rock types and structural chronology summarized in Table 9.7 may be generalized as follows:

Platform-type sequences:

1. metasedimentary sequences unconformably deposited upon older greenstones or sialic basement, for example, the Steep Rock Group (Wilks and Nisbet 1988; Thurston and Chivers 1990)

Keewatin-type sequences:

1. metasedimentary sequences of local extent within greenstone belts which lie unconformably upon older units but without intervening tectonic events, for example, Savant Lake group
2. conformable metasedimentary sequences above greenstones, for example, the White Partridge Bay group (Ayer 1987)

Timiskaming-type sequences:

1. metasedimentary sequences deposited unconformably upon older greenstones after an early deformation but before the late deformation, for example, Thompson Bay–Esox Lake sediments (Edwards 1983b; Smith 1990).

Plutonic Rocks

In this section, we describe the major granitoid units and discuss the relations between greenstone belts and granitoid units. The CWR includes an area with a granitoid basement to the greenstones, based on geochronologic (Davis and Jackson 1988) and stratigraphic (Wilks and Nisbet 1988) evidence. Granitic rocks have been subdivided into early gneissic and foliated tonalites, a later diorite-monzonite-granodiorite suite and a massive granodiorite to granite suite. There are minor bodies of a diorite-nepheline syenite suite and muscovite-bearing granitic rocks.

Granitic bodies in the Wabigoon Subprovince comprise large granitoid complexes surrounded by supracrustal rocks, smaller stocks and mafic to ultramafic intrusions that are commonly emplaced into supracrustal belts and minor

Table 9.5a. Mixed intermediate to felsic metavolcanic sequences: composition, rock types, textures and superpositionary relationships; a) western Wabigoon, b) eastern Wabigoon.

Group	Internal Units	Facies	Rock Type	Primary Structures	Geochemistry	Boundary Relationships	Major References	U-Pb Age (Ma)	Geochronology References
Kakagi Lake	Stephen Lake formation	distal and epiclastic	tuff; lapilli tuff; arenite; chert	bedded; cross-beds; graded beds; spherulites; soft-sediment deformation	calc-alkalic dacite	upper tectonic with Pipestone-Cameron deformation zone, lower conformable	Johns 1985	$2711.1^{+1.3}_{-1.2}$	Davis and Edwards 1982
	Cedartree Lake formation	distal and epiclastic	tuff; lapilli tuff; lapillistone; tuff breccia; intermediate and mafic dikes and sills	beds; graded beds; soft-sediment deformation	calc-alkalic dacite	upper and lower conformable			
	Emm Bay formation	proximal; distal-vent epiclastic	pyroclastic breccia tuff; breccia tuff; lapilli tuff; tuff; lapillistone; mafic, intermediate and felsic breccia; flows; intermediate and mafic intrusions	monolithic and heterolithic; massive; thick and thin unsorted beds; graded beds; flow banding and flames	calc-alkalic andesite; tholeiitic mafic flows	upper conformable, lower tectonic with East Kakagi Lake formation and unconformable with Snake Bay group		$2724.8^{+2.5}_{-2.3}$	
	East Kakagi Lake formation	distal; proximal; subvolcanic	tuff; lapilli tuff; intermediate and mafic intrusions	thick monolithic and heterolithic beds	calc-alkalic andesite	upper tectonic with Emm Bay formation, lower conformable			
	South Kakagi Lake formation	distal; proximal	tuff; lapilli tuff; arenite; chert	monolithic, thin graded beds; soft-sediment deformation	calc-alkalic andesite	upper conformable, lower disconformity with Katimiagamak group			
Dogpaw Lake volcanics			flows; tuff breccia; lapilli-tuff conglomerate	pillowed, massive, plagioclase-phyric, amygdaloidal flows; heterolithic; poorly sorted	interbedded calc-alkalic and tholeiitic rocks	upper tectonic with Wabigoon fault, lower interbedded with Rowan Lake volcanics	Johns 1987		
Cameron Lake volcanics	Upper Member	shallow water; proximal	pyroclastic breccia; tuff breccia; lapilli tuff; lapillistone; tuff; flows	plagioclase-phyric, pillowed, amygdaloidal, thick and thin beds with gradational contacts	calc-alkalic andesite	lower contact conformable with lower member	Johns 1986		
	Lower Member	proximal; distal	mafic flows; tuff breccia; tuff; wacke	massive and pillowed flows; monolithic pyroclastic rocks	interbedded calc-alkalic and tholeiitic volcanic rocks	lower conformable over Rowan Lake volcanics, upper conformable			

Table 9.5a. Mixed intermediate to felsic metavolcanic sequences: composition, rock types, textures and superpositionary relationships; a) western Wabigoon, b) eastern Wabigoon.

Group	Internal Units	Facies	Rock Type	Primary Structures	Geochemistry	Boundary Relationships	Major References	U-Pb Age (Ma)	Geochronology References
Hill Lake volcanics		distal; shallow water	Mafic flows; pillow breccia; hyaloclastite mafic tuff; lapilli tuff; tuff breccia	massive and pillowed; variolites; feldspar megaphyric; amygdaloidal; heterolithic; subrounded clasts	interbedded calc-alkalic and tholeiitic rocks	lower, conformable with Rowan Lake volcanics	Johns 1986		
Otterskin Lake volcanics			tuff breccia; tuff; lapilli tuff; pyroclastic breccia; flows; hyaloclastite	massive and pillowed; feldspar-phyric and sheared	interbedded calc-alkalic and tholeiitic rocks	lower, interbedded with Brooks Lake volcanics; upper tectonic with Pipestone–Cameron deformation zone	Edwards 1980		
Pincher Lake group	Manitumeig Lake formation		coarse, heterolithic, pyroclastic rocks; flows			lower gradational, upper intrusive	Blackburn 1982		
	Johnar Lake formation	deep water, shallow upwards	mafic flows; pyroclastic rocks	massive; amygdules	tholeiitic rocks	lower conformable			
	Noonan Lake formation		tuff; lapilli tuff; flows; tuff breccia		calc-alkalic rocks	lower conformable			
	Dog Lake formation		mafic flows	plagioclase-phyric	tholeiitic rocks	lower conformable			
	Garnet Bay formation		tuff breccia; tuff; felsic flow	lateral grading	calc-alkalic rocks	lower conformable			
	Stony Island formation		mafic flows; pyroclastic rocks	pillows; plagioclase-phyric	mixed tholeiitic rocks, calc-alkalic rocks	lower conformable			
Upper Manitou Lake group			intermediate and felsic pyroclastic rocks; mafic flows; sedimentary rocks		calc-alkalic rocks, minor tholeiitic rocks	lower disconformity over lower mafic group, upper conformable	Blackburn 1982		
South Sturgeon Lake volcanics	South Shore Cycle 5 formations	proximal vent; shallow water to subaerial; distal	interformational mixture of mafic flows and intermediate to felsic pyroclastic rocks	pillows; large amygdules; clast-supported fragmentals; many sedimentary features; fiamme	mixed tholeiitic and calc-alkalic rocks	lower hiatus conformable	Trowell 1983	$2717^{+2.7}_{-1.5}$	Davis and Trowell 1982

Table 9.5a. Mixed intermediate to felsic metavolcanic sequences: composition, rock types, textures and superpositionary relationships; a) western Wabigoon, b) eastern Wabigoon.

Group	Internal Units	Facies	Rock Type	Primary Structures	Geochemistry	Boundary Relationships	Major References	U-Pb Age (Ma)	Geochronology References
	Lyon Lake Cycle 4 formations	proximal, moderate water depth	layered mafic flows; felsic flows; mafic to intermediate flows; wacke; siltstone; conglomerated; sulphiditic mudstone	massive flows; autoclastic breccia; ripples; flaser convolute laminations	Mixed tholeiitic and calc-alkalic rocks	unknown	Trowell 1983	2736.0 ± 1.8	Davis et al. 1985
	Claw Lake Cycle 2 formations	shallow water, proximal	plagioclase-phyric mafic flows; intermediate to felsic fragmental hyaloclastite	coarse amygdules; pillows grade to thick selvages; phreatomagmatic brecciation of a felsic dome	mixed tholeiitic and calc-alkalic rocks	unknown	Trowell 1983	2735 ± 1.7	Davis et al. 1985
	Darkwater Lake Cycle 3 formations	deep-water shield volcanism, shallow up to stratovolcano, top subaerial	mafic flows; interflow sedimentary rocks; intermediate and intermediate to felsic fragmentals	pillows; fine and coarse amygdules; autoclastic breccia; hyaloclastite; pumice; monolithic; soft–sediment deformation; ore horizon	mixed tholeiitic and calc-alkalic rocks	unknown	Trowell 1983	$2736.3^{+9.3}_{-3.9}$	Davis et al. 1985
North Sturgeon Lake volcanics	Jumping Lake–Six Mile Lake Cycle 2 formations	proximal vent; shallowing water; subaerial?	mafic flows; intermediate to felsic fragmentals; wacke; conglomerate	massive; pillows; hyaloclastite; autoclastic breccia; thick selvages; amygdules; pyroclastic deposition; monolithic, angular clasts; accretionary lapilli; thick-graded beds	tholeiitic base, calc-alkalic top	structural	Trowell 1983		
North-east Arm volcanics	Richan Lake Cycle 1 formation		mafic flows; interflow wacke-siltstone			unknown	Trowell 1983		
	Beckington Lake East Cycle 2 formations		mafic flows; interflow sedimentary and felsic to intermediate pyroclastic rocks	massive; minor pillows; amygdules; porphyritic flows	tholeiitic rocks	unknown	Trowell 1983		

Table 9.5a. Mixed intermediate to felsic metavolcanic sequences: composition, rock types, textures and superpositionary relationships; **a)** western Wabigoon, **b)** eastern Wabigoon.

Group	Internal Units	Facies	Rock Type	Primary Structures	Geochemistry	Boundary Relationships	Major References	U-Pb Age (Ma)	Geochronology References
	Beckington Lake West Cycle 2 formations		mafic flows; interflow intermediate to felsic fragmentals; intermediate to felsic flows	massive; pillowed; fine amygdules; plagioclase-phyric; blue quartz eyes	tholeiitic and calc–alkalic rocks	unknown	Trowell 1983		
Handy volcanics		proximal to distal subaqueous	intermediate to felsic tuff; lapilli tuff; flows; mafic flows; wacke; conglomerate	plagioclase-phyric; massive; bedded	calc-alkalic rocks	upper, disconformity; lower, transitional	Bond 1980	$2745.2^{+1.9}_{-1.8}$	Davis and Trowell 1982
Lower Wabigoon volcanics			mixed mafic rocks and intermediate to felsic volcanic rocks		mixed tholeiitic and calc-alkalic rocks	lower unconformity; upper conformable		$2734^{+1.4}_{-1.2}$ $2742^{+4.0}_{-3.1}$	Davis et al. 1982
Clearwater Bay group	Clearwater formation	shallow submarine	mafic to felsic pyroclastic rocks; mafic to intermediate flows; sedimentary rocks	heterolithic, pillowed, amygdules	calc-alkalic rocks	conformable	Ayer 1990		
	White–shell River formation			wacke; siltstone; pelites		conformable			
	High Lake formation		mafic flows	massive; pillows; variolitic	tholeiitic rocks	conformable; unconformity with Crowduck Lake group			
	Zigzag Island formation		felsic pyroclastic rocks and flows		calc-alkalic rocks	conformable; unconformity with Crowduck Lake group			
Indian Bay group		shallow submarine	mafic flows; intermediate to felsic pyroclastic rocks	massive; pillowed; amygdaloidal; variolitic	tholeiitic and calc-alkalic rocks	conformable			
Andrew Bay group		shallow submarine	mafic to felsic volcanic rocks; sedimentary rocks; mafic flows		calc-alkalic and tholeiitic rocks	conformable			
Monument Bay group	Mason Lake formation	shallow submarine to subaerial	felsic to intermediate pyroclastic rocks	plagioclase-phyric	calc-alkalic rocks	conformable			

Table 9.5a. Mixed intermediate to felsic metavolcanic sequences: composition, rock types, textures and superpositionary relationships; a) western Wabigoon, b) eastern Wabigoon.

Group	Internal Units	Facies	Rock Type	Primary Structures	Geochemistry	Boundary Relationships	Major References	U-Pb Age (Ma)	Geochronology References
	Monument Bay formation	shallow submarine to subaerial	mafic to felsic pyroclastic rocks and flows	plagioclase-phyric and clinopyroxene-phyric	tholeiitic rocks; calc-alkalic rocks; shoshonitic rocks	conformable			
	Queen Island formation	submarine	sedimentary rocks; granodiorite; conglomerate	poorly bedded; moderate to well sorted; monolithic; well rounded					
	Cochrane Island formation		mafic to ultramafic flows; sedimentary rocks; felsic flows and pyroclastic rocks	variolitic; plagioclase megaphyric; massive; fine grained; thinly bedded sedimentary rocks	Mg-tholeiitic rocks; komatiitic rocks	conformable			
	Separation Point formation		mafic pyroclastic rocks and flows	clinopyroxene-phyric and plagioclase-phyric	calc-alkalic rocks	conformable			
	South Bay formation		mafic to intermediate flows and pyroclastic rocks	clinopyroxene-phyric and plagioclase-phyric	calc-alkalic rocks	conformable			
	Clipper Island formation		mafic to intermediate pyroclastic rocks and flows	clinopyroxene-phyric and plagioclase-phyric	calc-alkalic rocks	conformable			
	Redwater Bay formation		mafic flows	aphyric; pillowed to massive; variolitic	high-Mg tholeiitic rocks	conformable			
	Black Lake formation		mafic flows and pyroclastic rocks; sedimentary rocks; felsic pyroclastic rocks	massive; pillowed; aphyric; clinopyroxene-phyric	high-Mg tholeiitic rocks	conformable			
Royal Island group		distal to proximal	wacke; siltstone; mudstone; conglomerate; intermediate to felsic pyroclastic rocks	well bedded; graded; Bouma sequences		conformable			

Table 9.5a. Mixed intermediate to felsic metavolcanic sequences: composition, rock types, textures and superpositionary relationships; a) western Wabigoon, b) eastern Wabigoon.

Group	Internal Units	Facies	Rock Type	Primary Structures	Geochemistry	Boundary Relationships	Major References	U-Pb Age (Ma)	Geochronology References
Warclub group		distal to proximal	wacke; siltstone; mudstone; conglomerate; intermediate to felsic pyroclastic rocks	well bedded; graded; Bouma sequences		conformable			
Long Bay group	Skeet Island formation	shallow submarine; proximal to distal	mafic to intermediate flows; mafic to felsic pyroclastic rocks	plagioclase-phyric; clinopyroxene-phyric; pillows; amygdules	calc-alkalic rocks	conformable			
	Sunset Channel formation	shallow submarine; distal	intermediate to felsic tuff; wacke; lapilli tuff; tuff breccia	well bedded; plagioclase- and quartz-phyric	calc-alkalic rocks				
	Cliff Island formation		mafic to ultramafic flows; sedimentary rocks; felsic to intermediate siliceous pyroclastic rocks; wacke; siltstone	aphyric; pillowed to massive; variolitic; thinly bedded	Mg-tholeiitic rocks; komatiitic rocks	conformable			
	Gardner Island formation		interbedded felsic to intermediate tuff; lapilli tuff; tuff breccia; wacke; siltstone; mudstone	plagioclase- and quartz-phyric	calc-alkalic rocks	conformable			
	Poplar Island formation		wacke; siltstone; mudstone	thickly bedded					
Windigo Islands group	Drift Island formation	shallow submarine	intermediate to felsic pyroclastic rocks	plagioclase- and quartz-phyric	calc-alkalic rocks	conformable			
	Betty Island formation		wacke; sulphide siliceous iron formation			conformable			
	East Windigo Island formation		mafic pyroclastic rocks	clinopyroxene-phyric and plagioclase-phyric	calc-alkalic rocks				
	Burnt Island formation		wacke			conformable			

Table 9.5b. Mixed intermediate to felsic metavolcanic sequences: composition, rock types, textures and superpositionary relationships; **a)** western Wabigoon, **b)** eastern Wabigoon.

Group	Internal Units	Facies	Rock Type	Primary Structures	Geochemistry	Boundary Relationships	Major References	U-Pb Age (Ma)	Geochronology References
Elmhirst–Castle-wood domain	Elmhirst–Rickaby volcanic centre	shallow water to subaerial proximal vent	quartz-feld-spar porphyry; pyroclastic rocks	monolithic; spherulitic; massive flows	calc-alkalic rocks	lower comformable; upper not seen	Blackburn and Johns 1988		
	Gzowski–Oboshkegan volcanic centre	proximal vent; shallow water to subaerial	feldspar to quartz-feldspar porphyry	monolithic crumble breccia; thick, massive beds; little bedding	calc-alkalic rocks	upper conformable with mafic volcanic rocks; lower not seen			
Marshall–O'Sullivan lakes terrane	Marshall Lake volcanic centre	shallow water to subaerial; proximal to distal	mafic to inter-mediate flows and pyroclastic rocks; inter-mediate to felsic flows and pyro-clastic rocks	heterolithic, pumiceous, welded, spherulitic pillows; pillow breccia	calc-alkalic rocks, minor tholeiitic rocks	unknown	Amukun 1989		

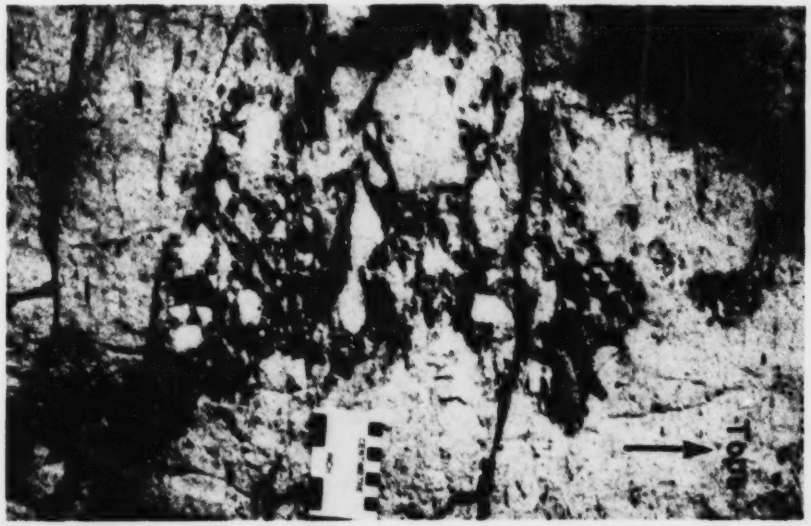


Figure 9.20. Gas escape structure in a monolithic block and ash flow, Berry River formation, Warclub group near Sioux Narrows. The narrowing of the brecciated area indicates tops (photograph *courtesy of* G. Johns).

Figure 9.21. Primary cross-bedding in mafic lithic and pyroxene crystal tuff, phreatomagmatic explosions forming a cinder cone. Cochrane Island formation, Monument Bay group, Monument Bay assemblage (photograph *courtesy of* G. Johns).

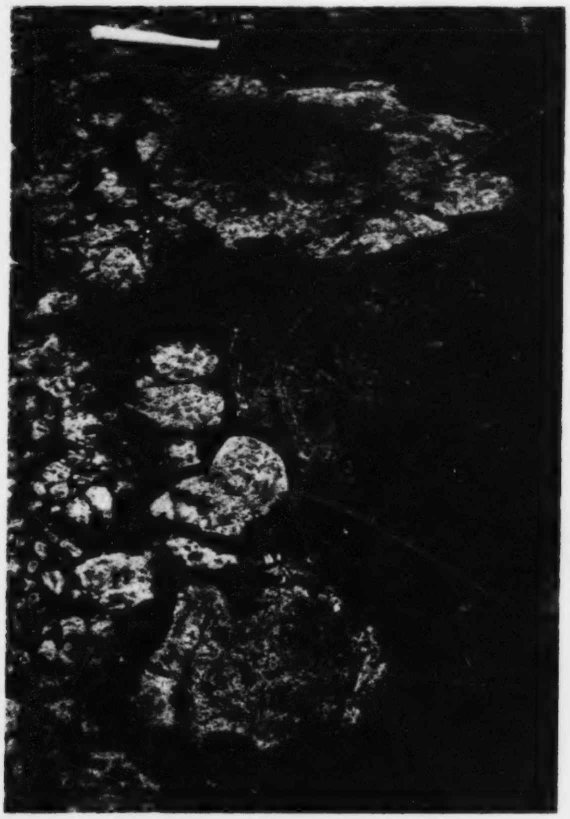


Figure 9.22. Reverse-graded ash flow with pumiceous clasts, Kawashegamuk volcanics, near Tabor Lake south of Dryden (photograph *courtesy of* C. Blackburn).

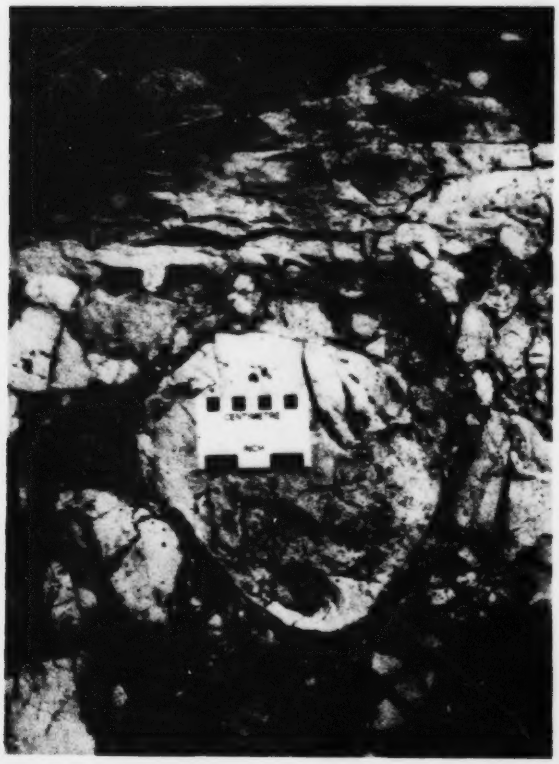


Figure 9.23. Rhyolitic tuff breccia with overlying tuff, Kawashegamuk volcanics, near Tabor Lake (*see* Figure 9.8) (photograph *courtesy of* C. Blackburn).

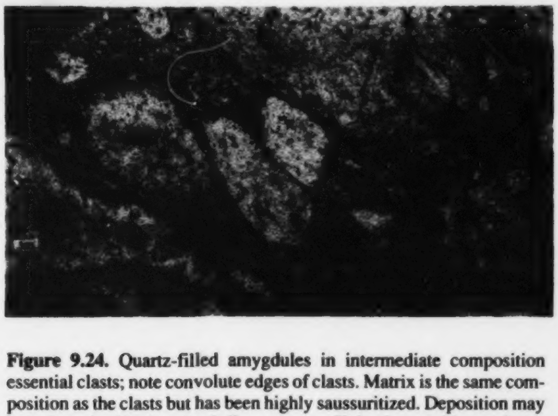


Figure 9.24. Quartz-filled amygdules in intermediate composition essential clasts; note convolute edges of clasts. Matrix is the same composition as the clasts but has been highly saussuritized. Deposition may have been by block and ash flow process; Kakagi assemblage (photograph *courtesy of* G. Johns).

alkalic intrusions. In the following sections, we describe elliptical to amoeboid batholiths that underlie about 50% of the WWR , whereas CWR is predominantly composed of domal gneissic and massive plutonic complexes and EWR is underlain by large batholithic complexes, separated by greenstone belts (*see* Figure 9.2).

Ultramafic to mafic plutons occur as large zoned or layered stocks, either marginal to, or emplaced into, granitoid batholiths, and as elongate, pretectonic sills within supracrustal sequences. Alkalic intrusions are rare; the Archean Sturgeon Narrows, Bell Lake, and Squaw Lake nepheline syenite intrusions and Proterozoic carbonatite diatremes at Chipman Lake are the only known bodies (*see* Figure 9.2). These alkalic intrusions are described by Sage (*see* Sage, this volume).

GRANITOID INTRUSIONS

Wabigoon Subprovince granitoid complexes can be divided into 3 types based on their age and geochemical relationships to the surrounding metavolcanic rocks:

1. Granitoid complexes range from ultramafic through to granitic in composition. These bodies consist of older sodium-rich granitoid marginal phases (with rare stocks which are mafic and/or ultramafic in composition) and a central region of younger potassium-rich granitoid units. These granitoid complexes are petrogenetically and geochronologically linked with volcanism (Davis et al. 1982; Edwards and Davis 1984; Davis and Edwards 1985; Shirey and Hanson 1986) and are here termed internal granitoid complexes. They are generally situated in the middle of greenstone belts with supracrustal strata younging

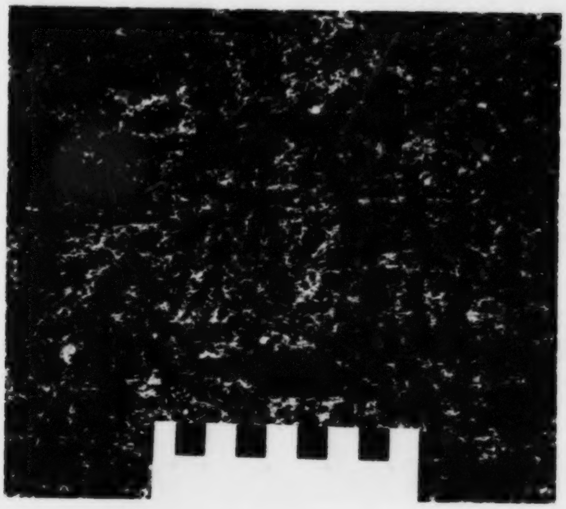


Figure 9.25. Clinopyroxene-phyric basaltic tuff from the Windigo Islands group, Lake of the Woods. The scale shown on the photograph is in centimetres. (photograph *courtesy of* J. Ayer).

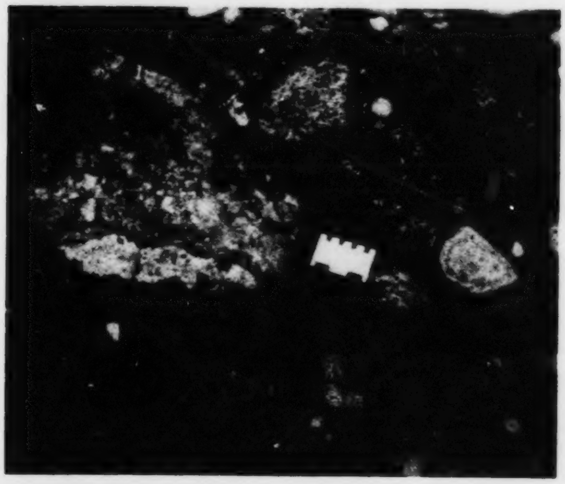


Figure 9.26. Tuff breccia derived from a submarine debris flow in the Andrew Bay group, Andrew Bay, Lake of the Woods. The variation in clast angularity, composition and the poor sorting is typical of this depositional process in the upper diverse sequences of the WWR (photograph *courtesy of* J. Ayer).

away from the complexes (e.g., *see* Figure 9.8). The age of emplacement of internal granitoid batholiths generally overlaps with calc-alkaline volcanism, for example, from 2732 to 2708 Ma in the WWR. Therefore, internal granitoid complexes do not form a basement to mafic, metavolcanic sequences.

2. Granitoid rocks form a basement complex underlying approximately 3 billion-year-old greenstones in the CWR. The Marmion Lake batholith, in the Steep Rock Lake area, dated at about 3 Ga (Davis and Jackson 1988), unconformably underlies the Steep Rock Group (Wilks and Nisbet 1988). The extent of the basement complex is impossible to constrain at present.
3. Posttectonic granitoid stocks both marginal and internal to supracrustal belts were emplaced later than the sub-jacent granitoid batholiths, either late in the tectonic evolution of the greenstone belts, where they may be unconformably overlain by graben-associated Timiskaming-type clastic metasedimentary rocks, or clearly postdating the final period of tectonism. They were emplaced between 2709 and 2685 Ma.

Basement Complex

Basement complexes in granite-greenstone subprovinces of the Superior Province are rare. The possibility of an unconformity beneath the approximately 3 billion-year-old Steep Rock Group, over the approximately 3 billion-year-old granitoids of the Marmion Lake batholith has been discussed for many years. Joliffe (1955) interpreted this contact to be an unconformity; Tanton (1927), Hicks (1950), and Schwerdtner et al. (1979) considered the Marmion Lake batholith to be intrusive into the Steep Rock Group, and Shklanka (1972) suggested a faulted contact. Recent work by Wilks and Nisbet (1988), and Stone et al. (in prep.) indicate that the contact is an unconformity.

Whereas, an unconformity beneath metasedimentary rocks in the Wabigoon Subprovince is not unique to the Atikokan area, the presence of platformal sedimentary rocks of sialic provenance (Wilks and Nisbet 1988; Thurston and Chivers 1990), and the approximately 3 billion-year-old age of the Marmion Lake batholith, make this a special case (*see* Figure 9.11). The Steep Rock Group is a quartz- and carbonate-rich platform sequence which may be in tectonic contact with the mafic metavolcanic Witch Bay Formation. Witch Bay Formation rocks and unnamed 2722 million-year-old metavolcanic rocks (Davis et al. 1989) are interdigitated in a zone along the Quetico Fault, raising some doubt about the age of the Witch Bay Formation. The relationship between the unnamed metavolcanic unit and the Witch Bay Formation is not entirely resolved. Most other geochronologic work in this area, on rocks of the Marmion Lake batholith, other stocks and gneisses external to the batholith and supracrustal rocks of the Lumby Lake greenstone belt, indicate a Mesoarchean age for these rocks. The Marmion Lake batholith has been dated at 3001 Ma, 2953 Ma and 3003±5 Ma (Davis and Jackson 1988). Two felsic tuffs and a quartz porphyry in the Lumby Lake belt are identical in age at 2999±1 Ma (Davis and Jackson 1988). Detrital zircon and a clast from the Steep Rock Group sedimentary rocks have been dated at 3001 Ma and 2994 Ma respectively, conformable with ages from the possible source rocks.

We suggest the Marmion Lake batholith and possible correlative units in the general area represent Mesoarchean basement in this area, but the extent of this basement is not known. The Marmion Lake batholith is cut by mafic dikes (Wilks and Nisbet 1988) and mafic dikes cut the *circa* 3 billion-year-old granitoids north of Armstrong (Davis et al. 1988). However, the presence of mafic dikes is not a universally applicable criterion for identification of *circa* 3 billion-year-old granitic rocks, as a unit of this sort on the

Table 9.6. Sedimentary sequences: facies associations, rock types, primary structures and superpositionary relationships.

Sequence (text figures)	References	Facies Associations	Units	Rock Types	Primary Structures	Lower Contact
Manitou Lakes group (Fig. 9.5)	Teal and Walker 1977; Teal 1979; Blackburn 1982	alluvial–fluvial	Uphill Lake formation	volcanic clast and polymictic conglomerates; pyroclastic rocks; sandstones and argillites	trough cross-bedded sandstones; lack of graded bedding in sandstones and conglomerates	Variable: in east, rubbly transition from mafic volcanic Sunshine Lake formation; in centre, sharp, conformable contact over Sunshine Lake formation; in west, indeterminate transition from Cane Lake subaerial pyroclastic rocks
Manitou Lakes group	ditto	resedimented	Mosher Bay formation	polymictic conglomerates; sandstones and argillites; minor magnetite-bearing iron formation	graded sandstones to argillites (Bouma sequences) interstratified with conglomerates; lack of cross-stratification.	conformable with underlying Uphill Lake formation; defined on basis of transition to graded sandstones and dominance of polymictic conglomerates
Manitou Lakes group	ditto	lacustrine	Surprise Lake member	laminated siltstone and argillite	none	minor member near base of Uphill Lake formation, intercalated in volcaniclastic conglomerates
Abram and Minnitaki groups (Fig. 9.6) Abram group	Turner and Walker 1973; Johnston 1972; Pettijohn 1935, 1943	alluvial	Ament Bay formation	polymictic, dominantly granitoid clast conglomerate; sandstones (arkosic) and pebbly sandstones	large-scale cross-stratification in sandstones; lack of graded bedding in conglomerates and sandstones	commonly faulted, but where observed, unconformably overlies Northern volcanic belt
Abram and Minnitaki groups (Fig. 9.6) Abram group	ditto	resedimented	Daredevil formation	tuff and lapilli tuff, and minor volcanogenic wackes and siltstones		conformably overlies Ament Bay formation
Abram and Minnitaki groups Abram group	ditto	resedimented	Little Vermilion formation	greywackes, >570 m thick	graded (Bouma sequences)	conformably overlies Daredevil formation
Abram and Minnitaki groups Minnitaki group	Walker and Pettijohn 1971; Pettijohn 1935, 1936; Johnston 1972	resedimented		wacke-siltstone, arkose and granitoid-clast conglomerate; minor magnetite-bearing iron formation. Preponderance of: conglomerates at east end, proximal facies; wacke-siltstones at west end, deep-water turbidites.	graded beds (Bouma sequences); lack of cross-stratification	unknown along much of belt, bounded by faults
Seine metasediments (Fig. 9.13)	Wood 1980	alluvial–fluvial, lacustrine?		polymictic conglomerate; interbedded conglomerate-sandstone; sandstone; siltstone; quartz-magnetite iron formation	poorly sorted conglomerates at base; trough cross-bedded sandstones; Wood (1980) suggests alluvial fan, passes up to braided fluvial environment and that thinly bedded iron formation at top indicates lacustrine conditions	unconformably overlies volcanic rocks (Keewatin), anorthosite (Bad Vermilion anorthosite) and tonalite (Bad Vermilion tonalite)

Table 9.6. Sedimentary sequences: facies associations, rock types, primary structures and superpositionary relationships.

Sequence (text figures)	References	Facies Associations	Units	Rock Types	Primary Structures	Lower Contact
Steep Rock Group (Fig. 9.11)	Shklanka 1972; Wilks and Nisbet 1988	platformal	Wagita Formation	fining-upward succession of conglomerate, sandstone, pebbly sandstone and pelite, up to 150 m thick	ortho- to paraconglomerate with clasts of subjacent rock types; sandstones well bedded, poorly sorted, but graded	unconformably overlies the Marmion Lake batholith
Steep Rock Group	ditto	platformal	Mosher Carbonate	stromatolitic carbonate (calcite, ankerite, dolomite; minor quartz, pyrite and kerogen); laminated, brecciated, up to 500 m thick	stromatolites: *Stratifera, Irregularia, Cryptozoon Walcotti*, Pseudo-fossil *Atikokania*, karst surfaces filled with Manganiferous Paint Rock Member	overlies Wagita Formation and may in part be lateral equivalent
Steep Rock Group	ditto	platformal	Joliffe Ore Zone Formation	Manganiferous Paint Rock Member: earth fragments of goethite, hematite, chert, quartz, matrix of clay minerals, 100 to 300 m thick. Overlain by—Goethite Member: brecciated lump ore of goethite (67%) and hematite (21%), plus quartz and kaolinite; also layers of goethite and chert; 50 to 100 m thick	fills karst depressions, contains Cretaceous spores suggesting secondary enrichment overlying dismal ashrock includes accretionary lapilli, suggesting shallow-water eruption	overlies Mosher Carbonate, and unconformably overlies Marmion Lake batholith; note that Goethite Member is overlain by the Dismal Ashrock Formation of the Steep Rock Group, an ultramafic pyroclastic rock.
Thompson Bay and Esox Lake sediments (Figs. 9.8 and 9.10)	Edwards 1983b; Thomson 1936a, 1936b; Smith and Stephenson 1988	alluvial-fluvial		arkosic wacke, rarely pebbly; siltstones, mudstones	graded sandstones with Bouma? sequences; cross-laminations; flame structures, load casts; rip-up clasts; clastic dikes.	unconformably overlies mafic metavolcanic rocks
Beardmore–Geraldton Belt (Fig. 9.14) Northern sedimentary belt	Devaney and Williams 1989; Devaney and Fralick 1985; and Barrett and Fralick 1989	fluvial	none named	polymictic conglomerate; pebbly sandstone	conglomerates massive to crudely bedded, pebbly sandstones plane laminated to cross-bedded	interpreted to lie with sheared unconformity on mafic volcanic rocks
Beardmore–Geraldton Belt Central sedimentary belt	ditto	fluvial to deltaic	Namewaminikan	south: mudstone, sandstone, iron formation central: sandstone, conglomerate, pebbly sandstone, mudstone north: conglomerate, pebbly sandstone	cross-bedded sandstone (braided river); herringbone cross-beds, muddy ripple drapes, trough cross-strata (paleoshoreline between deltaic to south and fluvial to north)	interpreted to lie with sheared unconformity on mafic volcanic rocks
Beardmore–Geraldton Belt Southern sedimentary belt	ditto	resedimented?		feldspathic wackes; thin-banded iron formation, conglomerate		interpreted to lie with sheared unconformity on mafic volcanic rocks
Stormy Lake group (Fig. 9.5)	McMaster 1978; Kresz 1984	redeposited to alluvial-fluvial?	none named	polymictic to volcaniclastic conglomerates, arkoses, predominant in west; sandstones (wackes) and interbedded iron formation in east	arkoses in west are trough cross-bedded; wackes in east are graded (Bouma sequences)	lies with angular unconformity on Wapageisi group in west; probably unconformable in east on Wapageisi group

Table 9.6. Sedimentary sequences: facies associations, rock types, primary structures and superpositionary relationships.

Sequence (text figures)	References	Facies Associations	Units	Rock Types	Primary Structures	Lower Contact
White Partridge Bay group (Fig. 9.9)	Ayer 1988a	redeposited to alluvial-fluvial	none named	sandstones, conglomerates, wackes	lower section contains fine-grained thinly bedded turbidites; upper part is medium-grained turbidites and conglomerates: coarsens upwards	
Crowduck Lake group (Fig. 9.9)	Davies 1965	alluvial?	none named	polymictic conglomerate, arkose, argillite, "agglomerate"		unconformably overlies mafic volcanic rocks of the Clearwater Bay group and the High Lake porphyry
Sturgeon–Quest–Post Lakes sediments (Fig. 9.3)	Trowell 1970	not defined	none named	wackes, arkoses, mudstones, siltstones, conglomerates	poorly stratified and/or graded	

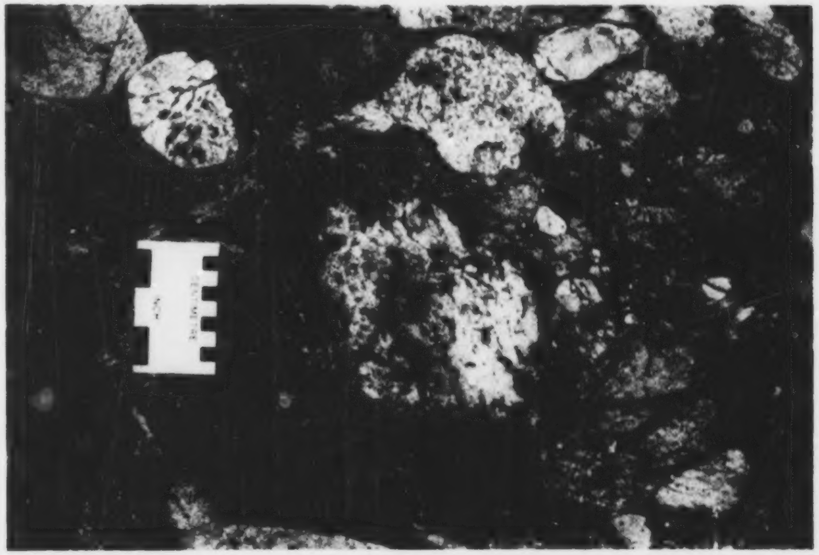


Figure 9.27. Large, red jasper clast above scale card in the alluvial fan-fluvial facies conglomerates of the Stormy Lake group (photograph *courtesy of* C. Blackburn).

Figure 9.28. Wacke-pelite beds with parallel and cross-laminations from the Warclub–Minitaki–Abram sediments, Eagle Lake east of Outlet Bay (photograph *courtesy of* C. Blackburn).

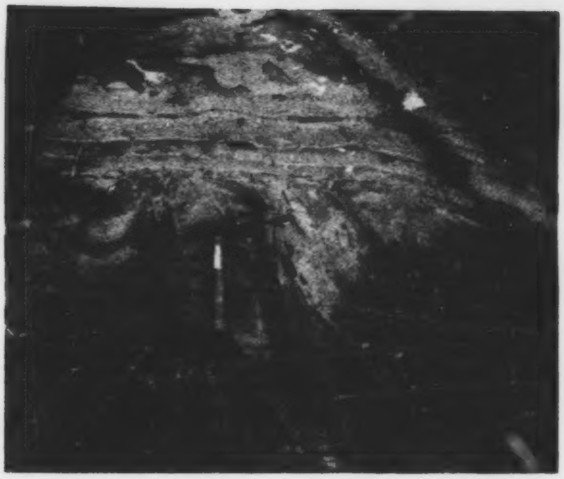


Figure 9.29. Medium-bedded normally graded wackes truncating large-scale slump folding in the underlying bed in the White Partridge Bay group, White Partridge Bay, Lake of the Woods (photograph *courtesy of* J. Ayer).

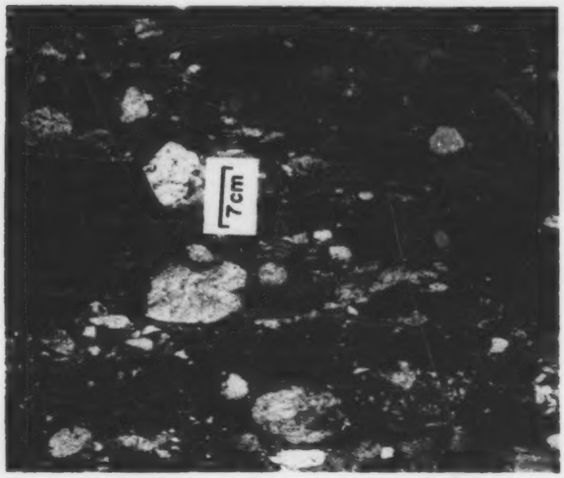


Figure 9.30. Conglomerate with heterolithic supracrustal and granitoid pebbles in the Crowduck Lake group, Crowduck Lake shore (photograph *courtesy of* J. Ayer).

Figure 9.31. Primary flow banding in a rhyolite flow in the Crowduck Lake group, Shoal Lake Road, south of Kenora (photograph *courtesy of* J. Ayer).

northeast shore of Lake Nipigon yielded a 2.7 billion-year-old age (R.H. Sutcliffe, Ontario Geological Survey, personal communication, 1988). Thurston and Davis (1985) have suggested that much of the CWR is underlain by a granitoid basement complex that may contain a number of Mesoarchean greenstone belt remnants. This suggestion was made on the basis of: the discordant, northeasterly trend of the central Wabigoon Subprovince; the local unconformable relationship at Atikokan; contrasting absolute ages of about 3.0 Ga in the Atikokan–Lumby Lake area and the 2.7 billion-year-old ages elsewhere in Wabigoon Subprovince; concentration of mafic and/or ultramafic, and syenitic and alkalic plutons around the margins of the CWR; and evidence for more than 1 metamorphism in the Lumby Lake belt (Davis and Jackson 1988), in contrast with a single event of an approximately 2.7 billion-year age elsewhere in the Wabigoon Subprovince. The extent, continuity and boundary relations of this suggested basement complex are not understood. Further support for Mesoarchean basement within Wabigoon Subprovince is found at Caribou Lake (*see* Figure 9.2), where tonalitic gneisses have been dated to be at least 3075±3 Ma (Davis et al. 1988). However, lack of geochronologic data throughout most of CWR, except for a preliminary age of 2750 Ma (D.W. Davis, Royal Ontario Museum, personal communication, 1990) from supracrustal rocks of the Obonga Lake greenstone belt, make the suggestion that it represents a large 3 billion-year-old basement complex, containing Mesoarchean greenstone belts, a hypothesis that needs much more testing.

Internal Batholithic Complexes

Neoarchean internal batholiths vary from tonalite and trondhjemite through granodiorite, quartz monzonite, and granite (Table 9.8), for example, Aulneau (Ziehlke 1974), Dryberry (Sanborn-Barrie 1991a), and Atikwa–Lawrence batholiths (Edwards and Davis 1984, 1986; Davis and Edwards 1985). Silica-poor phases, such as monzonite and syenodiorite are also present. Commonly, sodium-rich phases are older than potassium-rich ones, earlier phases tend to be foliated whereas later ones are massive.

Edwards and Davis (1984, 1986) and Davis and Edwards (1985) have demonstrated the Atikwa–Lawrence batholith contains 4 groups of rocks (units):

1. (Pyroxene)-biotite-amphibole diorite to amphibole-biotite tonalite (2733±1 Ma) accompanying and partially surrounding layered mafic intrusions (e.g., the $2733^{+1}_{-0.9}$ million-year-old Mulcahy Lake intrusion) occurs at the batholith-metavolcanic contact (Figure 9.32). This group is geochemically equivalent to andesitic volcanism in the surrounding greenstone belts.
2. Amphibole-biotite tonalite to granodiorite (2733±1 Ma), is geochemically equivalent to rhyolitic volcanism in the surrounding greenstone belt.

Table 9.7. Geochronology of the Wabigoon Subprovince.

Type 1: 2775 to 2718 Ma volcanic sequences and sodic plutonic rocks	Type 2: 2714 to 2696 Ma predominantly sedimentary sequences with minor calc-alkalic and alkalic volcanic units; intruded by syn- to late tectonic potassic plutons
1a: 2775 to 2745 Ma lower mafic sequences: dominantly submarine, high-Mg tholeiitic basalts; underlie Type 1b (diverse) sequences in the interior of the greenstone belt; generally in basal contact with and facing outward from, large synvolcanic batholiths – base of North Sturgeon Lake sequence, 2775±1 Ma – basalts underlying Handy Lake group, continuous? with above unit – Eagle Lake group, unconformably? underlying Lower Wabigoon sequence at Eagle Lake – Blanchard Lake group, underlying Type 1b (diverse?) sequence at Manitou lakes	turbidites: often in fault contact with underlying Type 2 volcanic sequences – Warclub and Minnitaki groups, <2713 ± 1 Ma and >2695 ±2 Ma in Minnitaki Lake area – Savant Lake group, may form eastward extension of Warclub group offset by Miniss River Fault – Central Sturgeon Lake sediments – Manitou Lakes group, Mosher Bay formation, <2706± 3 Ma and >2696 ±2 Ma – Coutchiching metasediments, Rainy Lake area, <2704± 3 Ma and >2692± 2 Ma
1b: 2745 to 2730 Ma upper diverse sequences: tholeiitic to calc-alkalic mafic-felsic sequences; coeval with underlying synvolcanic batholiths; found in middle to upper parts of assemblages, structurally underlying Type 2 turbidites (Warclub group) – Handy Lake group, 2745 ±2 Ma – upper part of North Sturgeon Lake sequence continuous with Handy Lake group? – lower 3 cycles of South Sturgeon assemblage, 2736 ± 1 Ma and 2735 ± 4 Ma for felsic units in the Lower Wabigoon volcanics intruded by the Atikwa Batholith at 2732 ±1 Ma – Fisher Lake volcanics continuous? with Upper Wabigoon volcanics—preliminary age of approximately 2732 Ma – Pincher Lake volcanics, continuous? with Lower Wabigoon volcanics	alluvial-fluvial sedimentary rocks: generally consist of locally derived conglomerate and sandstone, often underlying turbidite units and unconformably overlying Type 1b (lower mafic) units – Ament Bay formation (lower part), Abram group, <2698 ± 4 Ma—Savant Narrows formation (at Kashaweogama Lake), <2704 ± 1 Ma – Uphill Lake formation, Manitou Lakes group – Seine metasediments, Rainy Lake area, <2693 ± 2 Ma, >2686 ± 2 Ma; younger than sequences in the interior of the greenstone belt
Lower mafic sequences: dominantly submarine, Mg- to Fe-rich tholeiitic basalts; many of these sequences are marginal to the greenstone belt in basal contact with external granitoid plutons and facing toward the axis of the belt – Katimiagamak group, 2732 ±4 Ma – Snake Bay group appears to be continuous with Katimiagamak group – Wapageisi group along southern margin of belt between Pipestone and Stormy Lakes; continuous with Katimiagamak group, unconformably underlies Type 2b alluvial sedimentary rocks – Botham Bay group unconformably underlies Type 2b alluvial sedimentary rocks – Jutten group unconformably underlies Type 2b alluvial sedimentary rocks	volcanic sequences: generally intercalated with both types of sedimentary rocks; may be associated with subvolcanic porphyry intrusions – rhyolite near top of Kakagi Lake group (Stephen Lake), 2711 ±1 Ma – Berry River formation at base of Warclub group, 2714 ±6 Ma; felsic volcanic units and alkali basalts beneath and within sedimentary rocks of Manitou Lakes and Stormy Lake groups, <2699 ± 3 Ma – Daredevil formation in Abram group – felsic volcanic units in Minnitaki group, >2704 ± 3 Ma
1c: 2729 to 2725 Ma locally developed felsic volcanic centres – Phinney–Dash Lakes Complex, 2728 ± 1 Ma – eastern phase of the High Lake stock, grades into volcanic rocks 2727 ± 2 Ma – "Keewatin" volcanic units—Rainy Lake area, 2728–2725 Ma	
1d: 2725 to 2720 Ma differentiated mafic sills in coeval volcanic sequences – Boyer Lake group: preliminary ages of 2722± 5 Ma for a mafic sill and about 2720 Ma for a volcanic dacite – Kakagi Lake group: mafic sill, 2725 ±3 Ma; underlying Sabaskong batholith, 2724 ±2 Ma – Andrew Bay group: mafic sill 2724± 2 Ma; volcanic dacite, 2723 ± 2 Ma	
1e: 2720 to 2715 Ma relatively minor felsic units – upper cycle in South Sturgeon assemblage, 2718± 3 Ma – felsic porphyry in Vermilion Bay greenstone belt, 2716 ±2 Ma;	

Table 9.8. Synvolcanic batholithic complexes: rock types and tectonic relationships.

Batholith	Major References	Plutonic Events, Phases or Suites	Rock Types	Tectonic Relationship	U-Pb Age (Ma)	Geochronology References
Aulneau	Ziehlke 1974	early sodic event	foliated to geissic trondhjemites		2717^{+5}_{-3}	Davis and Edwards 1986
		first potassic event	trondhjemite to granodiorite to quartz monzonite			
		second potassic event	massive to foliated granodiorite and quartz monzonite		2710^{+4}_{-2}	Davis and Edwards 1986
Dryberry	Sanborn-Barrie 1990	older gneisses	gneissic tonalite, amphibolite, diorite and paragneiss		<2706 ± 2	Davis, R.O.M., personal communication, 1990
		sodic suite	tonalite and quartz diorite	synchronous with felsic volcanism, Vermilion Bay belt	2716 ± 2.1	Davis, R.O.M., personal communication, 1990
		potassic suite	granite		2663 ± 4	Davis, R.O.M., personal communication, 1990
Atikwa–Lawrence	Edwards and Davis 1984, 1986 Davis and Edwards 1985	group 1	diorite, quartz diorite to tonalite	synchronous with Mulcahy gabbro, dated at 2733 ± 1 Ma	2733 ± 1	Davis & Edwards 1985
		group 2	tonalite and trondhjemite to granodiorite	synchronous with Mulcahy gabbro	2733 ± 1	Davis and Edwards 1985
		group 3	trondhjemite and granodiorite to quartz monzonite		2719 (Atikwa) 2720 to 2725 (Lawrence)	Davis and Edwards 1985
		dikes	ultramafic to felsic dikes			
		Chuck Lake pluton	late granitic pluton		2718 inheritance	Davis and Edwards 1985
Sabaskong	Davis and Edwards 1986	gneiss	tonalite	early phase	2723 ± 2	Davis and Edwards 1986
		massive border phase	tonalite		2724 ± 2	Davis and Edwards 1982

3. (Amphibole)-biotite trondhjemite and granodiorite to quartz monzonite is the dominant phase, forming central parts of the batholith, and is dated at 2719 Ma (Atikwa lobe) with 2725 to 2720 million-year-old preliminary ages from 2 samples in the Lawrence Lobe.
4. Ultramafic to felsic dikes with abundant back-veining, net-veining, brecciation, and development of "pillows" are suggestive of magma mixing. The dikes represent injection of a late mafic magma into the lower felsic part of the batholith, suggestive of a late mafic underplating event (cf., Sutcliffe et al. 1989). In addition, Edwards and Davis (1984, 1986) and Davis and Edwards (1985) identify the late tectonic to posttectonic granitoid plutons (*see* Late Tectonic Stocks) as a distinct, later event.

The Atikwa–Lawrence batholith evolved over about 30 million years. (Edwards and Davis 1984, 1986; Davis and Edwards 1985). Field evidence indicates that units 1 and 2 were contemporaneous. Major central phases of the batholith (unit 3) are attributed to partial melting, at 2720 Ma, of basalt under conditions equivalent to the amphibolite-granulite transition caused either by crustal thickening during border zone magmatism, or by mafic underplating. Ultramafic to felsic dikes represent injection of basaltic or komatiitic melts into trondhjemite or granodiorite, resulting in fluxing of the adjacent granitoid rocks, partial mixing with more mafic magma and resultant injection higher in the crust as dikes or larger posttectonic quartz monzonite bodies, such as the Chuck Lake pluton.

Some synvolcanic batholiths have older gneissic units, for example, the Dryberry Batholith (Figure 9.33; Sanborn-Barrie 1991a). The Dryberry Batholith is composed of 30% older gneissic rocks of supracrustal and plutonic origin, into which homogeneous intrusive rocks forming 70% of the batholith were emplaced. Sanborn-Barrie (1991a) has recognized 3 suites of granitoid rocks on the basis of field relationships, petrology, and major- and trace-element geochemistry. A high Al_2O_3 -type tonalite suite is oldest, and concentric plutons of the granodiorite suite intrude the tonalite. Rocks of the granite suite occur mainly in the east and also intrude the tonalite suite but predate the granodiorite suite.

Sanborn-Barrie (1991a) points out that the heavy REE depletion of the tonalite suite is consistent with partial melting of a parental basaltic source at mantle depths, but high-Al_2O_3 trondhjemites might also be generated by fractional crystallization of a hornblende gabbro magma at crustal levels. Field evidence to support this latter mechanism includes the presence of mafic to intermediate rocks related to the tonalitic suite by a process of fractionation from a tonalite suite parent magma. REE profiles for granodiorite suite rocks show steep patterns compared with other Archean granodiorites.

On this basis, Sanborn-Barrie (1991a) reasons that simple partial melting of crustal rocks to produce the granodiorite is not consistent with their REE geochemistry and suggests two-stage partial melting of a crustal source to account for overall-depleted REEs. If greywacke is the source, Sanborn-Barrie (1991a) calculates that a 35% partial melt of wacke would account for production of the granite suite, leaving the residue for second stage partial melting to produce granodiorites.

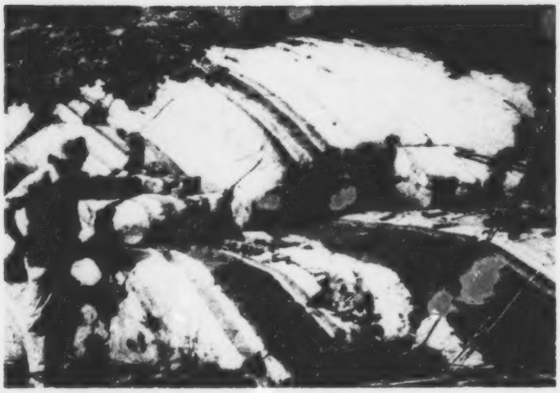


Figure 9.32. Layered norite-anorthosite of the Mulcahy Lake gabbro, part of the Atikwa–Lawrence batholith. Outcrop is south of Mulcahy Lake (photograph *courtesy of* C. Blackburn).

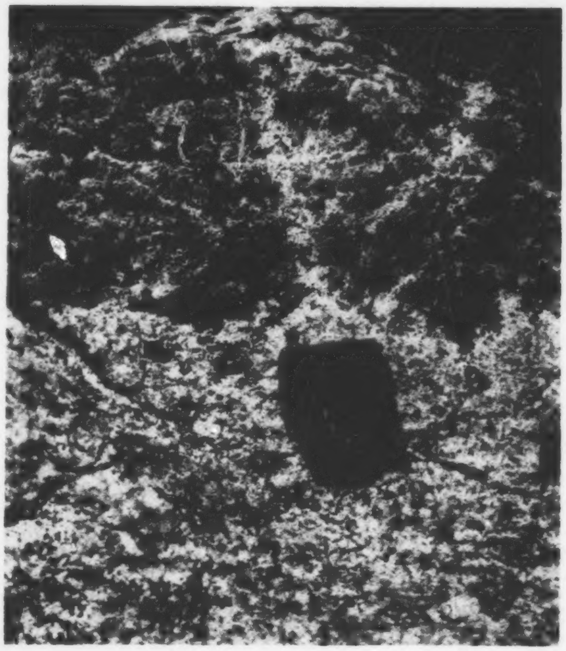


Figure 9.33. Dryberry Batholith, coarse anorthositic gabbro intruded by granodiorite to granite pegmatite (photograph *courtesy of* G. Johns).

The elliptical Aulneau Batholith (*see* Figures 9.2 and 9.10; Ziehlke 1974) is composed of 3 major intrusive pulses: 1) foliated to gneissic trondhjemites in the western half of the batholith dated at 2717 Ma; 2) later, potassium-rich trondhjemite-granodiorite to quartz monzonite, comprising discrete intrusive phases through the northwestern quadrant of the batholith; and 3) massive to foliated granodiorite and quartz monzonite, representing a second stage of potassium-rich magmatism at 2717 Ma that occupies most of the eastern half of the batholith. Ziehlke (1974) mapped 24 separate phases in the Aulneau Batholith, but the relative age relationships of a number of other minor phases were not apparent. A gabbro to diorite body at the western margin of the batholith predates the earliest trondhjemites and contains numerous, large, xenoliths of the adjacent volcanic rocks of the Lake of the Woods greenstone belt. The absolute age of the batholith together with its elliptical shape and little evidence of dioritic border phases, suggests that it is more akin to the late tectonic stocks described in a later section.

The Irene–Eltrut Lakes and Sowden–Wabikimi Lakes batholithic complexes (Percival 1983; *see* Figure 9.2) occupy most of the CWR. In both complexes, Sage et al. (1975) recognized 2 major suites: foliated tonalite to granodiorite and massive granite to granodiorite. Percival (1983) has subdivided these, introducing a gneissic tonalite to granodiorite suite and late gabbroic bodies.

Gneissic Complexes of Central Wabigoon Region

Many of the gneissic plutonic rocks of CWR consist of tonalitic gneisses some with supracrustal remnants consisting of amphibolitic screens or inclusion trains. Scanty geochronologic data discussed in Table 9.7 suggest that in future this unit may be divided into early, possibly Mesoarchean units and Neoarchean syntectonic complexes. Schwerdtner et al. (1979) pointed out that many of the bodies have a domal structure now considered (Schwerdtner 1990) to be incompatible with either magmatic or solid-state diapirism and that the structure may be due to cross folding in a subhorizontal stress field. The domal gneisses are commonly separated from the adjacent greenstone belts by late crescentic intrusions of massive granodiorite or quartz monzonite. An example of these bodies is the Rainy Lake batholith (*see* Figures 9.2, 9.10 and 9.13) which has been shown by Sutcliffe and Fawcett (1980) to be composed of 4 distinct elements: 1) septa of amphibolite-facies supracrustal rocks within the oldest of the granitoid suites; 2) an early suite of highly foliated to gneissic tonalite-granodiorite; 3) foliated to massive diorite to monzonite to granodiorite of the crescentic Jackfish Lake–Weller Lake pluton (and a smaller pluton at Spencer Lake); and 4) a late suite of foliated to massive biotite granite and pegmatite.

Sutcliffe and Fawcett (1980) suggested, in conformity with the models of Hanson and Goldich (1972) and Arth and Hanson (1975), that the early tonalite to granodiorite suite was derived by partial melting of a basaltic precursor at mantle depths (*see also* Longstaffe (1977) on models for genesis of tonalites at Footprint Lake in the western part of the batholith). The Jackfish Lake–Weller Lake pluton, which contains abundant mafic clots and aggregates, is considered by Sutcliffe et al. (1990) to be representative of the sanukitoid suite, formed as a juvenile mantle-derived melt with variation in the suite due to hornblende fractionation during the later stages of Archean granitoid plutonism. The third suite (element) is considered by Sutcliffe and Fawcett (1980) to have been derived by partial melting of the early tonalite-granodiorite suite. A preliminary zircon U-Pb age of 2696±1 Ma (D.W. Davis, Royal Ontario Museum, personal communication, 1990) has been obtained from monzodioritic rocks of the Jackfish Lake–Weller Lake pluton. This age indicates that large parts of the batholith are considerably younger than much of the volcanism in adjacent greenstone belts.

The Irene–Eltrut Lakes batholithic complex (*see* Figure 9.2) is composed of foliated tonalite-granodiorite, intruded by later massive granitic plutons. The White Otter Lake batholith is by far the largest of these plutons. It is composed of massive granite to granodiorite, devoid of xenoliths, except in the border regions where fragments of earlier phases are common. Other plutons (Sage et al. 1975) include: the Norway Lake pluton, zoned from syenodiorite and monzonite to syenite; the Entwine Lake intrusion, composed of an early augite-syenodiorite phase and later biotite-augite diorite and porphyritic monzonite phases; and the Pekagoning and Smirch Lake plutons, both massive monzonite to diorite.

The Sowden–Wabikimi Lakes batholithic complex (*see* Figure 9.2), separated from the Irene–Eltrut Lakes batholithic complex by supracrustal screens, is similarly divisible into early magmatic foliated and also gneissic, tonalite to granodiorite and late magmatic massive to weakly foliated granodiorite, quartz monzonite and granite. Most of the demonstrably late magmatic plutons are in the southern half of the complex and include (Sage et al. 1975): the Cecil Lake pluton, composed of massive quartz monzonite to granodiorite, and comparable to the White Otter Lake batholith, except that gneissic xenoliths are abundant; the Basket Lake batholith, composed of tonalite to granodiorite; the McCausland pluton, composed of granodiorite to quartz monzonite; and the Stedman Lake pluton, composed of tonalite to granodiorite. Farther north are (Goodwin and Howat 1977) the Metionga Lake, Seiss Lake, Harmon Lake and North Brennan Lake plutons. In the centre of the batholithic complex, a gabbro-pyroxenite body (Empire Lake gabbro) and a diorite-monzonite-syenite-gabbro complex (Colwill Lake pluton) have been identified (Percival 1983). Other gabbro-pyroxenite bodies occur along the eastern margin of the complex (e.g., Lac des Iles, Outlet Bay; Sutcliffe and Smith 1988).

Late Tectonic Stocks

Late tectonic to posttectonic stocks that intrude the supracrustal belts (Figure 9.34) range from quartz diorite through quartz monzonite to granite, with a concomitant range in silica-poor varieties from diorite through monzonite to syenite. The stocks thus lack the sodium-rich phases characteristic of the synvolcanic batholiths (cf., Dryberry Batholith, with over 50% sodium-rich phases). Similar bodies also form ovoid or crescentic plutons intruding the gneissic complexes of central Wabigoon. More than 30 stocks (Table 9.9) assigned to this unit (unit 15, *see* Bedrock Geology of Ontario, Maps 2541 to 2545, map case), based upon field

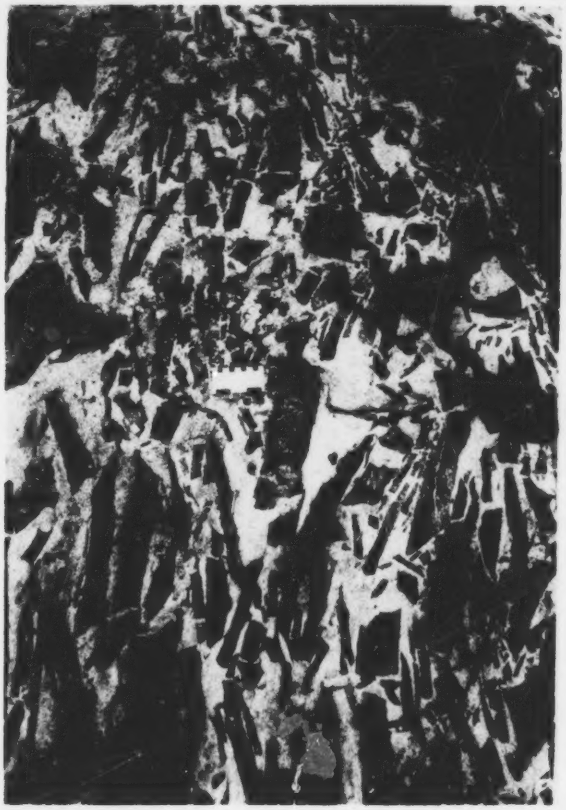


Figure 9.34. Intrusion breccia of granodiorite (Berry Lake stock), intruding the Snake Bay group (photograph *courtesy of* G. Johns).

relations, are distributed throughout the supracrustal belts of Wabigoon Subprovince. The majority are in WWR.

The stocks vary from homogeneous, to inhomogeneous, to zoned, and many show evidence of multiple episodes of magma emplacement. Zoned intrusions (e.g., Flora Lake, Black Hawk, Ryckman Lake and Ottertail Lake) are concentrated in the WWR, and are characterized by granodioritic cores, and monzodioritic rims (Heimlich 1965; Davies 1973b; Blackburn 1973, 1976a). Inhomogeneous stocks show a similar variation; the Taylor Lake stock (Blackburn 1981; Pichette 1976) varies from an early monzodiorite through monzonite and quartz monzonite, to granodioritic phases. Other inhomogeneous stocks include the Burditt Lake stock (Blackburn 1976b), the High Lake stock (Davies 1965; Pedora 1976) and the Canoe Lake stock (Davies 1965; Campbell 1973). There is considerable variation between homogeneous stocks; however, most are granodioritic to quartz monzonitic in composition.

Many of these stocks, especially those of intermediate composition, are enriched in both incompatible and compatible elements (Shirey and Hanson 1984; Stern et al. 1989; Sutcliffe et al. 1990). Shirey and Hanson (1984) demonstrated that this unusual chemical property is characteristic of the sanukitoid suite found in some Phanerozoic arc settings. This suite is considered to originate by partial melting of enriched mantle. More felsic examples of the late plutonic suite, which do not show enrichment in incompatible elements, may be derived from fractional crystallization of these melts.

MAFIC TO ULTRAMAFIC INTRUSIONS

Ultramafic to mafic plutons form large zoned or layered stocks, either marginal to, or emplaced into, granitoid batholiths and as elongate, pretectonic sills within supracrustal sequences. The types of mafic and ultramafic intrusions in the WWR are (*modified from* Sutcliffe *in* Macdonald and Cherry 1988):

1. sills synchronous with mafic volcanism in greenstone belts
2. syntectonic intrusions at the margins of major batholiths, coeval with dacitic volcanism and early stages of felsic plutonism such as those associated with the Atikwa–Lawrence batholith (e.g., Mulcahy gabbro [Davis and Edwards 1985; Sutcliffe 1984a], Nabish Lake Gabbro, Mile Lake–Trap Lake Intrusion, bodies in the Denmark Lake–Populus Lake area [Davies and Watowich 1958] and the Mitchell Lake intrusion)
3. syntectonic to posttectonic intrusions coeval with late felsic plutonism, for example, Lac des Iles intrusive complex (Sweeny and Sutcliffe 1986) and the Entwine Lake intrusion (Davies 1965)
4. serpentinized ultramafic intrusions such as the Chrome Lake–Puddy Lake intrusion (Whittaker 1980)

Mafic and ultramafic intrusions of the Wabigoon Subprovince have been summarized in Table 9.10.

Intrusions Related to Mafic Volcanism

Examples of intrusions considered synchronous with mafic volcanism are the Katimiagamak Lake sills dated at $2731^{+4}_{-2.9}$ Ma (Davis and Edwards 1985), the Bad Vermilion anorthosite (Phinney et al. 1988) and numerous sills within the Bigstone Bay group and the Rowan Lake volcanics. A later pulse of tholeiitic magmatism is represented by the Kakagi Lake sills dated at $2728^{+2.5}_{-2.3}$ Ma (Davis and Edwards 1982). The Bad Vermilion Lake anorthosite is the most thoroughly studied example; being a 100 km^2 lens within WWR greenstones near Mine Centre (*see* Figure 9.2; Harris 1974; Wood et al. 1980a, 1980b; Ashwal et al. 1983). Uranium-lead data from a nearby gabbroic unit and the adjacent felsic volcanic and intrusive rocks indicate an age of about 2728 Ma (Davis et al. 1989). The Bad Vermilion Lake complex consists of anorthosite, leucogabbro and gabbro. Coarse, equant, 1 to 30 cm plagioclase grains in a finer mafic groundmass characterize the anorthosite. Layering is present but is not well developed. Gabbroic rocks contain local concentrations of iron-titanium oxides and/or apatite, but no chromite. Ashwal et al. (1983) pointed out that the bulk composition of the complex is equivalent to aluminous basalt, but REE abundances are unlike those of typical Archean high-Al_2O_3 basalts. This difficulty was addressed by Phinney et al. (1988), who suggested fractionation of plagioclase in crustal magma chambers and the addition of lesser amounts of evolved basaltic liquid.

Table 9.9. Synvolcanic to posttectonic stocks: structure, rock types and tectonic relationships.

Stock	Major References	Structure	Phases	Rock Types	Tectonic Relations	U-Pb Age (Ma)	Geochronology References
High Lake	Davies 1965 Pedora 1976	inhomogeneous	early	porphyritic granodiorite to trondhjemite	synvolcanic	2727 ±2	Davis and Smith 1991
			late	tonalite and granodiorite	late tectonic	2711 ±2	Davis and Smith 1991
Canoe Lake	Davies 1965 Campbell 1973				late to posttectonic	2709 ±1	Davis and Smith 1991
Viola Lake	Ayer 1988a	homogeneous		granodiorite	late to posttectonic		
Falcon Island	Ayer 1988b	inhomogeneous	1	syenite to alkalic syenite	late tectonic		
			2	monzonite	late tectonic		
			3	pyroxenite to monzogabbro	late tectonic?		
Stephen Lake	Davies and Morin 1976	homogeneous or heterogeneous		granodiorite? or quartz diorite?	posttectonic	2699 ±2	Davis and Edwards 1986
Flora Lake	Davies 1973b Heimlich 1965	zoned	early (rim)	monzodiorite	late to posttectonic		
			middle (mantle) monzonite	late to posttectonic			
			late (core) granodiorite (granite?)	late to posttectonic			
Hope Lake	Heimlich 1966	inhomogeneous	early	monzonite	late to posttectonic		
			late	granite	late to posttectonic		
Nolan Lake	Kaye 1974a, 1974b	heterogeneous		monzonite, monzodiorite	late to posttectonic		
				quartz monzonite	late tectonic		
Froghead Bay	Moorhouse 1941	homogeneous		"granite"	late tectonic?		
Phinney and Dash lakes	Edwards 1983a	inhomogeneous	early to late	syenite and leucocratic trondhjemite	synvolcanic	2728 ±1	Davis and Edwards 1982
Burditt Lake	Blackburn 1976b	inhomogeneous	north	grey granodiorite	late tectonic		
			south	pink granodiorite	late tectonic		
Black Hawk	Blackburn 1976b	zoned	rim	monzonite	late to posttectonic		
			core	granodiorite	late to posttectonic		
Ottertail Lake	Harris 1974 Poulsen 1984b	zoned	rim	monzodiorite to diorite	posttectonic	2686^{+5}_{-3}	Davis et al. 1989
			core	granodiorite to granite	posttectonic		
Taylor Lake	Blackburn 1981, 1982; Pichette 1976	inhomogeneous	early	monzodiorite	posttectonic	2695 ±4	Davis et al. 1982
			middle	monzonite, quartz monzonite	posttectonic		
			late	granodiorite			
Beidelman Bay	Trowell 1983	inhomogeneous		trondhjemite, granodiorite, quartz diorite and quartz-feldspar porphyry	synvolcanic	2734	Davis and Trowell 1982

Table 9.10. Mafic to ultramafic intrusions: form, rock types and tectonic relationships.

Intrusion	Major References	Intrusion Type (see text)	Rock Types	Tectonic Relations	U-Pb Age (Ma)	Geochronology References
Bigstone Bay sills	Ayer et al. 1987	1 (differentiated)	peridotite, pyroxenite, melagabbro, gabbro, leucogabbro	synchronous with mafic volcanism		
Kakagi Lake sills		1 (differentiated)	peridotite, pyroxenite, gabbro, leucogabbro	synchronous with mafic volcanism	2728 ±2	Davis and Edwards 1986
Katimiagamak Lake sills	Edwards 1983b	1	gabbro, leucogabbro, peridotite	synchronous with mafic volcanism	2732^{+4}_{-3}	Davis and Edwards 1986
Mulcahy Lake	Sutcliffe 1984a, 1984b, 1984c Morrison et al. 1986	2 (layered)	norite, gabbro, gabbronorite, pyroxenite, anorthosite, troctolite	syntectonic and coeval with Atikwa–Lawrence batholith	2733 ±1	Morrison et al. 1985
Bad Vermilion Lake and Grassy Portage sills	Harris 1974 Wood et al. 1980	1	anorthosite, leucogabbro, gabbro	synchronous with mafic volcanism	2727 ±2	Davis et al. 1989
Lac des Iles intrusive complex and Tib gabbro	Sutcliffe et al. 1989 pyroxenite	3	gabbro, gabbronorite,	syntectonic to posttectonic coeval with late felsic plutonism	2692^{+4}_{-2}	Davis, unpublished data
Chrome Lake–Puddy Lake sill	Whittaker 1980	4	dunite, peridotite, pyroxenite	Alpine-type intrusions tectonically emplaced in a crystalline state		

Intrusions Related to Late Volcanism and Early Plutonism

The Mulcahy Lake intrusion (2733.2 Ma) is typical of the synvolcanic intrusions (Davies and Watowich 1958; Sutcliffe 1984a; Morrison et al. 1986); it intrudes the Populus Lake volcanics and is in turn intruded by the Atikwa–Lawrence batholith (*see* Figure 9.2). The intrusion is 63 km^2 and is subdivided into a lower (2 to 2.5 km), middle (2.5 to 2.8 km) and an upper zone (1 to 1.5 km) defined by mapping and mineral chemistry. The lower and middle zones represent distinct pulses of magma which crystallized from the base upward. The upper zone was considered by Morrison et al. (1986) as a region which cooled from the roof downward, whereas Sutcliffe and Smith (1985) suggested that the roof zone is composed of mafic to intermediate hornfels intruded by sills of peridotite to granite in composition. Major phases consist of norites, gabbros and gabbronorites with smaller amounts of pyroxenites, anorthosites and troctolites, commonly as discrete marker horizons. Rhythmic layering is well developed in lower and middle zone cumulates. Marginal parts of the lower and middle zones, 150 to 300 m wide, are characterized by higher concentrations of apatite and magnetite (Morrison et al. 1986). Three zones of the intrusion have elevated values for Cu, Ni and Cr. The elevated values are found: a) in orthopyroxene cumulates 1 km above the base of the lower zone; b) at the lower zone-middle zone boundary; and c) at the troctolite unit in the middle zone (Macdonald and Cherry 1988).

Intrusions Related to Late Plutonism

The Lac des Iles intrusive complex (Pye 1968; Watkinson and Dunning 1979; Sutcliffe and Smith 1988; Sutcliffe et al. 1989) is the best known of the late mafic-ultramafic intrusions in the EWR. The complex is a 30 km^2 body (*see* Figure 9.2) and is 2691±3 million years old, contemporaneous (based on field relations) with late granitoids and cutting early 2729±2 million-year-old gneissic tonalite. The complex is one of several intrusions which have broadly similar rock types, relative ages and metallogenetic potential. It forms a circular structure about 30 km in diameter north of Thunder Bay (*see* Figure 9.2). Other intrusions in the group are the Tib Gabbro (Smith and Sutcliffe 1986), Demars Lake (Kaye 1969), Dog River, Taman Lake (Sutcliffe 1988) and Legris Lake intrusions (Pye 1968). Sutcliffe (*in* Macdonald and Cherry 1988) considered the intrusions to be similar to the Mesozoic, Alaskan zoned ultramafic complexes (Irvine 1967; Taylor 1967) based upon: 1) the occurrence of marginal zones of hornblende-rich ultramafic rocks; 2) the type of rock type variation caused by the interplay of fractionation and addition of new magma; and 3) the association of ultramafic rocks with gabbroic units which are not closely related to each other by simple fractionation. The complex is host to concentrations of platinum group element (PGE) and gold mineralization associated with disseminated copper and nickel sulphides.

The complex consists of several circular to elliptical mafic to ultramafic intrusive centres (Gupta and Sutcliffe 1990). The 3 distinct magma sequences are hornblende gabbro, gabbro and gabbronorite and ultramafic rocks. The ultramafic sequences were emplaced late and are concentrated in the north end of the complex. Cyclic modal layering is only developed in the northern, ultramafic part of the complex and chromite is present as minor disseminated grains and is locally concentrated as layers up to 1 cm thick. The major PGE mineralization occurs along the interface between gabbro and gabbronorite and associated pyroxene cumulate in the southern part of the complex. The mineralization is localized by mixing the PGE and sulphide-rich gabbronorite and pyroxene cumulates with water-rich gabbro and has been redistributed by late magmatic volatile activity which generated mineralized gabbroic pegmatites and breccia zones (Sutcliffe et al. 1989).

Tholeiitic basaltic magma of picritic and high-Al_2O_3 affinity are postulated as parents to the ultramafic and gabbroic parts of the complex respectively. The magmatic evolution of the complex involved multiple pulses of previously fractionated magma, some of which underwent subsequent *in-situ* fractionation (Sutcliffe et al. 1989).

The fact that the Lac des Iles intrusive complex is cogenetic with surrounding granitic rocks suggests that the magma, a juvenile mantle melt (Sutcliffe et al. 1990), is a possible source for mafic magmas which may be involved in mafic underplating late in Archean tectonism.

Serpentinized Ultramafic Intrusions

The representative of this group is the Chrome Lake–Puddy Lake intrusion at the northern contact of the Obonga Lake greenstone belt. The Chrome Lake–Puddy Lake intrusion is serpentinized peridotite with irregularly distributed lenses of pyroxenite and dunite. The body is bounded to the north by quartzofeldspathic metasedimentary gneiss, with which it has a sheared contact. To the south, the ultramafic rocks are also bounded by a shear zone from rocks of the Obonga Lake greenstone belt.

DISCUSSION

Types 1 and 2 are the most abundant mafic and/or ultramafic intrusions in the Wabigoon Subprovince. The geochronologic data indicate that at least 2 broad periods of mafic intrusion occurred in the Wabigoon Subprovince. Mafic intrusion occurred: 1) synchronous with volcanism and early plutonism at about 2.73 Ga, for example, Bad Vermilion and Mulcahy; and 2) in the late stages of granitic plutonism at about 2.69 Ga, for example, Lac des Iles

The mafic to ultramafic intrusions of the WWR are largely tholeiitic, but identification of parental magma compositions has been hampered by the lack of chilled margins and by the contamination of marginal and dike phases. Microprobe analyses of primary phases constrain magma composition; the magnesian tholeiite suite is represented by the Lac des Iles intrusive complex (Sutcliffe et al. 1989) and the iron tholeiite suite by intrusions such as the Mulcahy gabbro (Sutcliffe 1984a). Anorthositic intrusions represent plagioclase cumulates with minor amounts of interstitial liquids (Phinney et al. 1988).

Primary mineralogy is commonly preserved in the large intrusions, therefore crystallization sequences have been established for some intrusions. Most intrusions have the sequence olivine, pyroxene and plagioclase. The first formed pyroxene is variable in composition, depending probably on silica activity (Sutcliffe *in* Macdonald and Cherry 1988), with orthopyroxene preceding clinopyroxene at high-silica activity. In the Mulcahy Lake and Kakagi Lake sills, the order of crystallization of cumulus minerals is olivine, orthopyroxene, clinopyroxene and plagioclase; whereas at Lac des Iles it is olivine, clinopyroxene, orthopyroxene and plagioclase. In anorthositic intrusions, crystallization sequences are olivine, plagioclase and pyroxene (Phinney et al. 1988).

STRUCTURAL GEOLOGY

Introduction

A complex and variable structural style in the granite-greenstone Wabigoon Subprovince distinguishes it from the adjacent Quetico and English River metasedimentary-dominated subprovinces (Wilson 1949; Gill 1949). Interior parts of the Wabigoon Subprovince consist typically of large elliptical to irregular granitoid domes, separating intervening, subordinate cuspate wedges of little deformed (Figure 9.35) supracrustal rocks (*see* Figure 9.2). In contrast to the interior parts of the subprovince, the northern and southern margins are marked by supracrustal sequences occurring as linear belts paralleling the subprovince boundaries and are relatively uninfluenced by granitoids.

In the interior of the subprovince, belts and wedge-shaped masses of supracrustal rocks commonly face and dip away from the domes. Rise of the granitoid domes into an extensive layer of supracrustal rocks may in part have been responsible for producing the apparent synformal nature of greenstone belts.

Long, sinuous shear zones commonly follow the central axes of the supracrustal belts. These shear zones exhibit complex histories of brittle and ductile deformation with variable kinematics. For example, the Monument Bay, Pipestone–Cameron, and Manitou Straits fault system (*see* Figures 9.9, 9.10 and 9.40) may be traced a distance of 250 km from the central part of Lake of the Woods in the west, to Dinorwic Lake in the east.

The northern and southern margins of the subprovince display a linear structural grain manifested by repetitive volcanic and sedimentary sequences in which stratigraphic facing determined from local top directions can be inward, outward or consistent, for example, near Sioux Lookout in the Abram and Minnitaki lakes belts (*see* Figure 9.12) and Beardmore–Geraldton Belt (*see* Figure 9.14; *see also* Williams, this volume, Figure 10.4). The pattern of repeated units, in association with structural evidence of layer-parallel shearing, strongly supports the hypothesis that supracrustal assemblages have been tectonically stacked. Geochronologic data confirm the juxtaposition of old rocks above younger (Davis et al. 1988). Contacts between and within these belts of volcanic and metasedimentary rocks

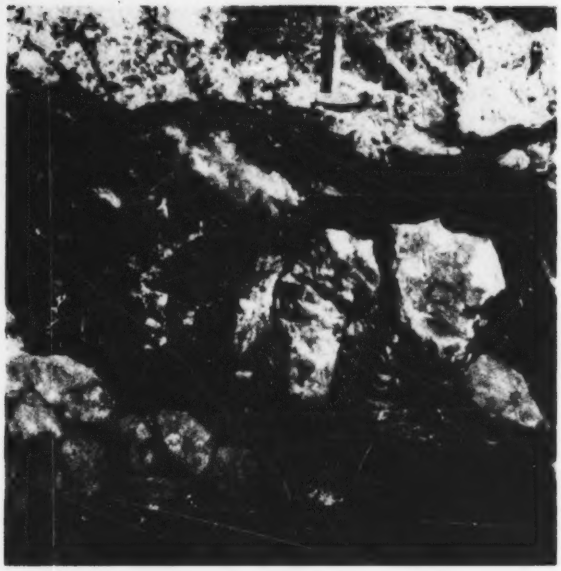


Figure 9.35. Columnar jointed massive basalt flow from the Bigstone Bay group, Hades Islands, Lake of the Woods (photograph *courtesy of* J. Ayer).

are commonly faulted. Some of these faults may be traced many tens of kilometres, for example, the Little Vermilion fault and the Wabigoon fault along the northern boundary in the Dryden–Sioux Lookout area (*see* Figure 9.12), and the Paint Lake deformation zone (*see* Figure 9.14) that forms the boundary between normal granite-greenstone terrane to the north and the Beardmore–Geraldton Belt (Williams 1990; *see also* Williams, this volume, Figure 10.4).

We next describe the style, scale, timing and distribution of folds, shear zones and faults in the subprovince using examples from areas where detailed structural analysis has been carried out. Two major structural regimes appear to have been operative: an initial aggregation of supracrustal assemblages of similar, or contrasting age, to produce both simple and complex greenstone belts; and interaction of the Wabigoon Subprovince with its neighbours, producing marginal tectonic features that do not affect the internal parts of the Wabigoon Subprovince. A proposed correlation of deformation events established by a number of workers for each of these regimes, is shown in Table 9.11. The data reveal a general evolution of structural style, but the synchroneity of an individual style cannot be assumed.

Deformation Structures

Several folding and shearing deformations affect the greenstone supracrustal rocks of the Wabigoon Subprovince. The structural style of these progressive deformations is different within the body of the Wabigoon Subprovince, from that seen at its northern and southern margins. The distribution of folds and shear zones and their assignment to specific relative deformation phases is shown in Figures 9.36a, 9.36b, 9.36c and 9.36d.

Major patterns of deformation within greenstone belts appear to consist of an early, layer-parallel deformation that is subsequently reworked by upright fold systems and major shear zones.

FOLD AND SHEAR ZONE DEVELOPMENT WITHIN GREENSTONE BELTS

Within the internal parts of the subprovince, the axial traces of tight folds (*see* Figure 9.36a-d), determined from regional changes in stratigraphic facing in supracrustal assemblages, are for the most part parallel to greenstone belt trends and batholith margins (e.g., Manitou anticline; *see* Figure 9.8). Later, more open folds may have axial traces that are belt-parallel, or trending at high angles to batholith contacts; some of these may be attributed to batholithic emplacement (e.g., Bigstone Bay anticline; *see* Figure 9.9), but others postdate at least some magmatism. Much of the interior part of the subprovince has undergone 2 periods of deformation: an initial one that produced a pervasive layer-parallel schistosity or was premetamorphic; and a later folding that deformed this planar fabric. The trend and plunge of small-scale fold axes reflect local conditions and are strongly dependent on proximity to large domal batholiths. On a mesoscopic scale, refolding may be demonstrated, particularly adjacent to shear zones and faults. Kink bands and other brittle-ductile shear fabrics (Figure 9.37; e.g., Blackburn 1976a; Edwards and Stauffer 1988) are not distributed homogeneously but are spatially associated with major shear zones and faults that lie along greenstone belts or transect the subprovince.

Structural analysis carried out by Smith (1990) and Smith and Stephenson (1988) around Vista Lake allowed them to demonstrate that an early phase of north-plunging folding was refolded by easterly trending upright folds. An example of the complexity that may be present is given by an area just west of Vista Lake where Edwards and Stauffer (1990) recognized 7 stages of deformation (*see* Table 9.11, Figure 9.36a). The earliest is a set of isoclinal, north-trending D_1 folds of unknown, possibly steep plunge, interpreted from opposing stratigraphic-facing directions in mafic volcanic rocks. The D_2 deformation structures consist of steeply plunging upright folds; this is again interpreted on the basis of opposing facing directions. In neither of the first 2 deformations have the fold axes been recorded. The D_2 fold axial traces are deformed by a northeasterly plunging D_3 fold. Subsequently, small-scale D_4 folds and an associated east-striking, steeply south-dipping cleavage overprint the first 3 structural features. Later deformations are only locally recognized and are mainly associated with the Manitou Stretch–Pipestone Lake shear zone.

In the Savant Lake region, Sanborn-Barrie (1989) recognized 2 stages of folding prior to motion on the Kashaweogama Lake fault (*see* Table 9.11). Axial traces of D_1 folds are variable, trending from northerly to easterly but are refolded by northeast-trending and plunging D_2 folds (*see* Figure 9.36c).

The combined structural data described above from throughout the subprovince (*see* Table 9.11) illustrates that

Table 9.11. Correlation of structural development in the Wabigoon Subprovince. Map areas representative of the core of the subprovince and its northern and southern marginal zones are grouped. Progressive structural development has been grouped into 3, represented by: early, flat-lying structures; upright folding; and strike–slip shearing.

Area; Reference	Early Structures: Imbrication, Shallow-plunging Folds; (Early structures)	Upright, Steeply Plunging Folds (Belt-parallel structures)	Late Structures: Brittle-Ductile Shearing (Post-accretion convergence)
North Boundary Zone			
High Lake; Smith 1987	large-scale shallow N-plunging, NW-trending folds, no fabric	E-trending, steeply E-plunging folds, strong schistosity	dextral sense, N side up
Shoal Lake, Duport; Smith 1987; Davies and Smith 1988	open to isoclinal N-trending and plunging folds, perhaps associated with pluton emplacement	E-trending open folds	dextral shearing associated with Crowduck-Rush Bay deformation zone (D.Z.) and Shoal Lake D.Z.; kinematics: NW side up, then S side up
Monument Bay; Morrice 1989	recumbent to shallow-dipping early schistosity, N-plunging mineral lineation	E-NE-trending, steeply W-plunging folds, strong schistosity	dextral shearing on Monument Bay D.Z.
Lake of the Woods; Ayer et al. 1987; Sanborn–Barrie 1991b	large-scale folds paralleling NE- to E-trending schistosity, SW plunging	steeply E-plunging folds, steep E- and W-plunging lineations	dextral shearing, N side up on Crowduck-Witch Bay D.Z.
Chisholm Island; Ayer and Buck 1989	layer parallel schistosity with upright shallow E-plunging isoclinal folds	tight, E-trending, steeply W-plunging folds; open, N-trending syncline related to pluton emplacement	dextral shearing along Pipestone-Cameron D.Z., S side up
Dryden; Blackburn 1979	no fabric or early structures recognized	tight, steeply, west-plunging Z-folds	open, steeply plunging Z-folds associated with dextral shearing along Wabigoon Fault
Sandybeach Lake; Chorlton 1991	recumbent isoclinal folds and early flat-lying schistosity, NE-trending folds followed by emplacement of small plutons	NE–trending, belt-parallel folds, metamorphism, sinistral shearing, with SE side up	dextral shearing along Wabigoon fault, SE side up
Wabigoon Core Zone			
Cameron-Rowan lakes; Melling 1989	no early fabric or structures recognized	E-NE-trending, SW-plunging, W-facing major fold, Shingwak antiform	dextral shearing along Pipestone–Cameron–Manitou D.Z.
Flint–Cameron lakes; Osadetz 1979; Buck 1988	layer-parallel schistosity, shallow N-plunging, N-trending folds	E-NE-trending, steeply E-plunging folds	dextral shearing along Pipestone–Cameron–Manitou D.Z.
Pipestone Lake; Edwards and Stauffer 1990	N-plunging isoclinal folds	E-plunging and trending isoclinal folds; N-NE–plunging open folds; E-trending cleavage overprinting all folds	dextral shearing along Pipestone–Cameron–Manitou D.Z.
Vista Lake; Smith 1990	N-plunging folds	E-trending folds with steep E plunge; steeply N-plunging folds	dextral shearing along Pipestone–Cameron–Manitou D.Z.
Savant Lake; Sanborn-Barrie 1990	N- and E-trending, steeply plunging isoclinal folds	NE-trending, asymmetric folds with steep E plunge	dextral shearing along Kashaweogama Lake fault, N side up
South Boundary Zone			
Rainy Lake; Poulsen et al. 1980; Davis et al. 1989	imbrication and thrusting, layer-parallel shearing and small-scale folding	upright, subprovince-parallel folding	dextral shearing along both Quetico and Rainy Lake-Seine River faults
Beardmore–Geraldton Belt; Kehlenbeck 1986; Devaney and Williams 1989; Williams 1990	imbrication, thrusting and layer-parallel shearing of volcano-sedimentary couplets, small-scale folds of variable plunge, down-dip lineations	shallow-plunging mesoscopic belt-parallel folding	emplacement of Croll Lake stock at 2692 Ma during late stages of dextral shearing, shallowly W-plunging Z folds, faulting along steeply dipping, E-trending layering

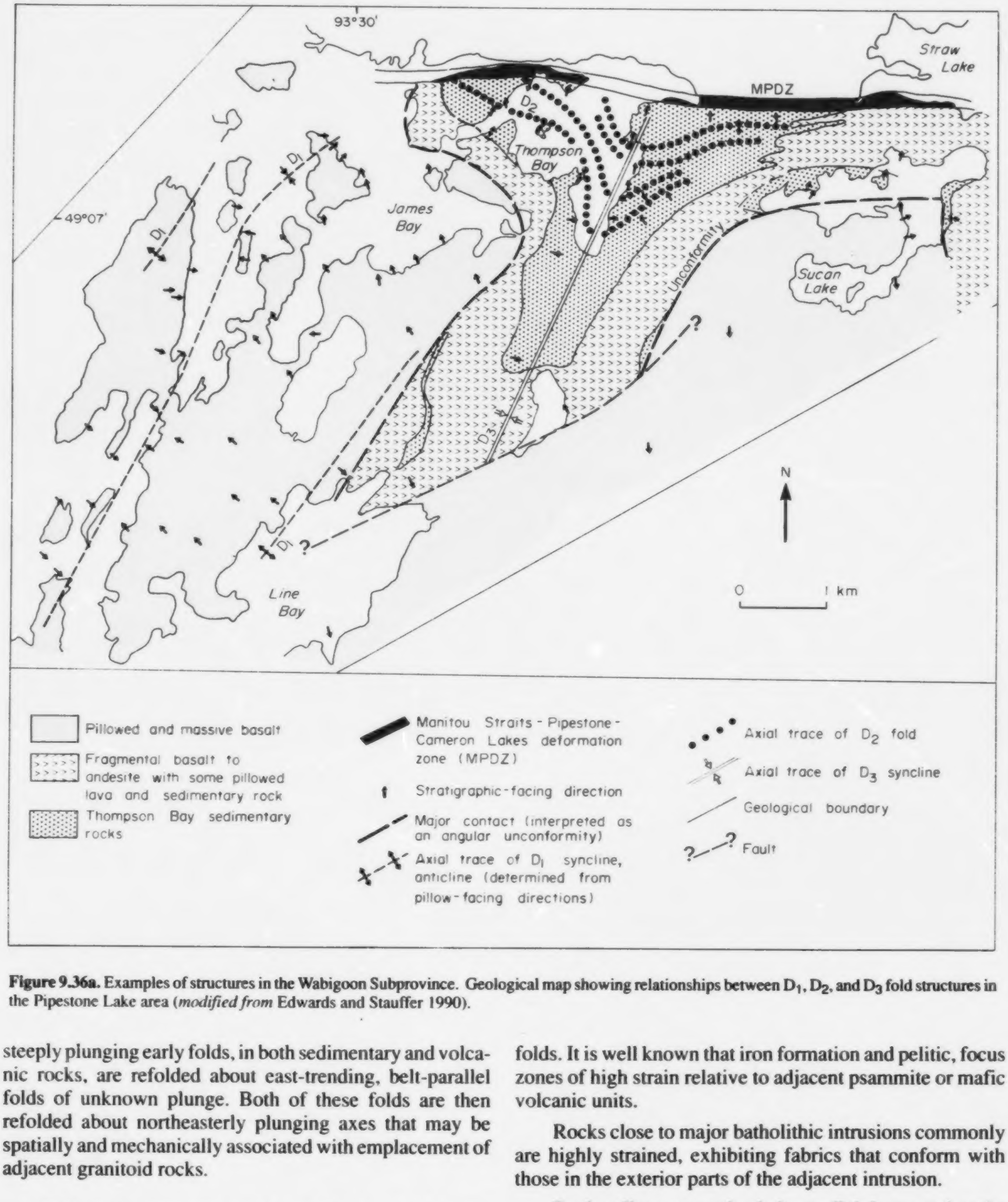

93°30'
Straw Lake
MPDZ
D2
Thompson Bay
D1
D1
D1
D1
49°07'
James Bay
Unconformity
Sucan Lake
D3
?
?
N
0
1 km
Line Bay
Pillowed and massive basalt
Fragmental basalt to andesite with some pillowed lava and sedimentary rock
Thompson Bay sedimentary rocks
Manitou Straits - Pipestone - Cameron Lakes deformation zone (MPDZ)
Stratigraphic-facing direction
Major contact (interpreted as an angular unconformity)
Axial trace of D1 syncline, anticline (determined from pillow-facing directions)
Axial trace of D2 fold
Axial trace of D3 syncline
Geological boundary
Fault

Figure 9.36a. Examples of structures in the Wabigoon Subprovince. Geological map showing relationships between D_1, D_2, and D_3 fold structures in the Pipestone Lake area (*modified from* Edwards and Stauffer 1990).

steeply plunging early folds, in both sedimentary and volcanic rocks, are refolded about east-trending, belt-parallel folds of unknown plunge. Both of these folds are then refolded about northeasterly plunging axes that may be spatially and mechanically associated with emplacement of adjacent granitoid rocks.

Table 9.11 documents much variation in the intensity, geometry and attitude of folds and shear zones within greenstone belts. This variation may in part be due to local and regional mean rock ductility and scale of ductility contrasts, and in part to proximity to major batholith intrusions and belt-parallel shear zones. For example, it is established, in many greenstone belts, that deformation processes are recorded by sedimentary rocks more faithfully than their volcanic counterparts, both as internal shear strain and as folds. It is well known that iron formation and pelitic, focus zones of high strain relative to adjacent psammite or mafic volcanic units.

Rocks close to major batholithic intrusions commonly are highly strained, exhibiting fabrics that conform with those in the exterior parts of the adjacent intrusion.

Rocks adjacent to major, belt-parallel shears such as the Manitou Straits–Pipestone–Cameron Lakes deformation zone, or the Wabigoon fault are strained for several hundred metres on either side of the trace of the structure (Figures 9.38 and 9.39). With each of these major shears, is a number of genetically associated fold systems and minor or splay shear zones.

An understanding of local structural geology within a greenstone belt is therefore dependent on a number of local

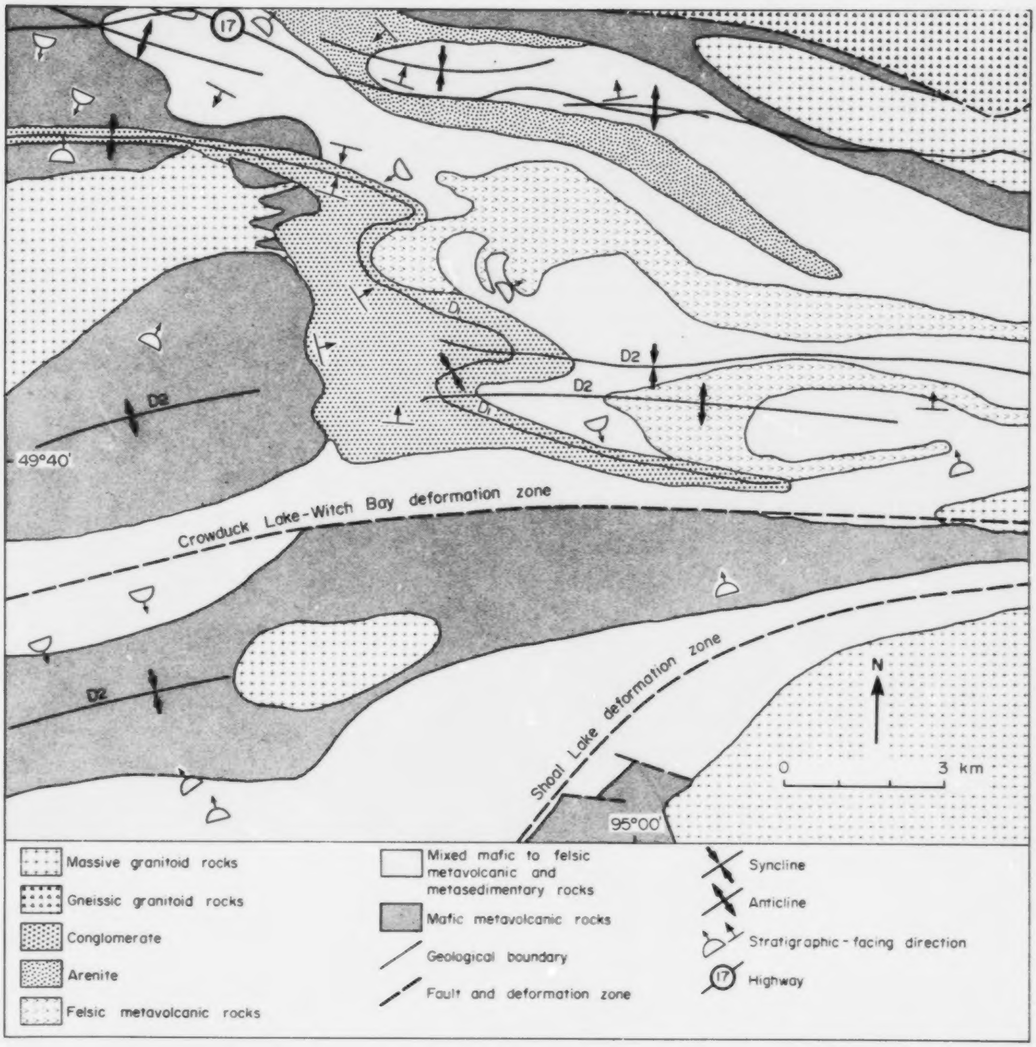

17
D1
D2
D2
D2
D1
D2
D2
49°40'
Crowduck Lake-Witch Bay deformation zone
Shoal Lake deformation zone
N
0
3 km
95°00'
Massive granitoid rocks
Gneissic granitoid rocks
Conglomerate
Arenite
Felsic metavolcanic rocks
Mixed mafic to felsic metavolcanic and metasedimentary rocks
Mafic metavolcanic rocks
Geological boundary
Fault and deformation zone
Syncline
Anticline
Stratigraphic - facing direction
17
Highway

Figure 9.36b. Examples of structures in the Wabigoon Subprovince. Geological map showing relationships between D_1, and D_2 fold structures in the High Lake–Rush Bay area, Lake of the Woods (*modified from* Ayer 1990).

parameters. With the benefit of our present understanding, we have not recognized a throughgoing fabric development, anisotropy or direction.

Fabric Development in Plutonic Complexes

Granitoid plutons and batholithic complexes surround the greenstone belts and have played an integral part in their deformational history. For many years, the granitoid complexes were considered to be diapiric in origin (Schwerdtner et al. 1979). Schwerdtner (1990) reinterpreted the patterns of lineations and foliations within and around the Ash Bay dome and Western Rainy Lake complex granitoid masses in the southern part of the Wabigoon Subprovince, recognizing an early recumbent fabric that was subsequently domed by upright folding. Schwerdtner (1990) suggested that the early fabrics were formed during subhorizontal shearing. An early fabric of the above type may have been recognized by Morrice (1989) on the north side of the Sabaskong batholith at Monument Bay, where the shallow dips seen in supracrustal rocks are coplanar with the contact of and foliations within the batholith to the south.

Central Shear Zones

Long, sinuous zones of shearing and faulting characterize the central, or axial parts of supracrustal greenstone belts, for example, Monument Bay, Pipestone–Cameron and Manitou Straits faults in the cental part of the western part of the subprovince, the Crowduck Lake–Witch Bay deformation zone in northern Lake of the Woods (*see* Figure 9.9) and the Sturgeon Lake fault (*see* Figure 9.3). Numerous splays extend from these shear zones, in some cases into wedge-shaped masses of supracrustal rocks developed

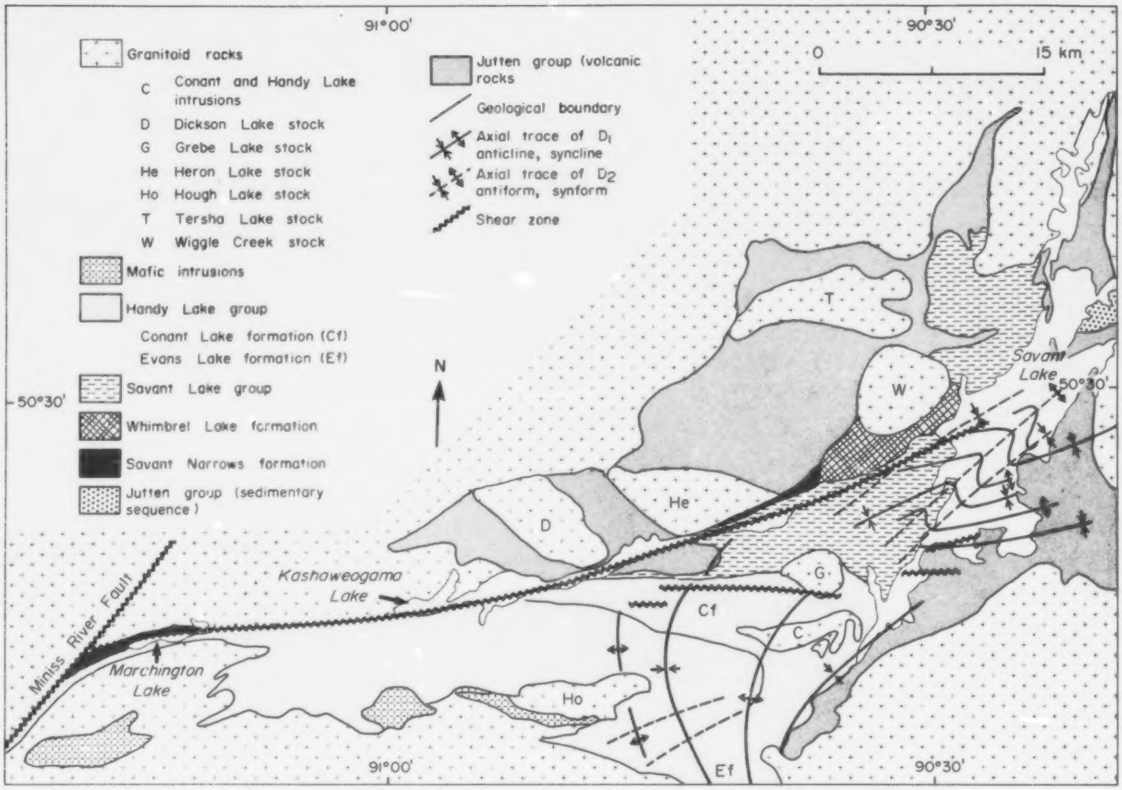

91°00'
90°30'
0
15 km
Granitoid rocks
C Conant and Handy Lake intrusions
D Dickson Lake stock
G Grebe Lake stock
He Heron Lake stock
Ho Hough Lake stock
T Tersha Lake stock
W Wiggle Creek stock
Mafic intrusions
Handy Lake group
Conant Lake formation (Cf)
Evans Lake formation (Ef)
Savant Lake group
Whimbrel Lake formation
Savant Narrows formation
Jutten group (sedimentary sequence)
Jutten group (volcanic rocks
Geological boundary
Axial trace of D1 anticline, syncline
Axial trace of D2 antiform, synform
Shear zone
N
50°30'
Savant Lake
Kashaweogama Lake
Marchington Lake
Miniss River Fault
T
W
He
D
G
Cf
C
Ho
Ef

Figure 9.36c. Examples of structures in the Wabigoon Subprovince. Regional-scale D_1 and D_2 folds, and shear zones in the Savant Lake area (*modified from* Sanborn-Barrie 1990).

peripherally to the main belts, such as the zone of deformation along the central part of the Burditt Lake belt, extending south from the Kakagi–Rowan Lakes greenstone belt (*see* Figure 9.10). In some cases the splays occur within the main belt, such as the Shoal Lake deformation zone at Shoal Lake (*see* Figure 9.9; Smith and Stephenson 1988) and the Monte-Cristo shear zone (*see* Figure 9.10).

Central shear zones vary in width, fabric and strain intensity, both within and between shears. The end of the Manitou Straits fault and shear system is marked by a zone of highly schistose, carbonate to sericite schist several tens of metres wide along its southeast side, whereas a less intensely deformed zone up to 3 km wide, to the northwest of this, occurs along much of its length. Horse-tail splays mark the termination of many shear systems. At the northeast and southwest ends of the Manitou Straits fault, deformation is distributed amongst a set of horse tails at Dinorwic Lake and Lower Manitou Lake respectively.

Movement sense and degree on central shear zones has been difficult to ascertain. Despite the juxtaposition of contrasting lithologic sequences across the shear zones, correlation across the zones has rarely been possible. Structural analyses (e.g., Melling 1989; Buck 1988; Ayer and Buck 1989; Smith 1990) of tectonic fabrics such as mineral lineations and slickensides allowed determination of the orientation of movement lines. Similarly, deflection of schistosity, and observed asymmetry of small-scale C-S fabrics (Berthé et al. 1979), of minor folds, served as kinematic indicators. Unfortunately, these indicators commonly record the latest movement sense only, which may not have been the major component, and do not indicate amount of movement.

Steeply plunging mineral and stretching lineations on the planar fabrics developed in association with the central shear zones, coupled with lack of evidence of major offset of distinctive units on either side of the shears, suggests that the major component of movement on them has been dip-slip, rather than strike-slip.

FOLD AND SHEAR ZONE DEVELOPMENT AT SUBPROVINCE MARGINS

Fold patterns within the northern and southern marginal zones of the Wabigoon Subprovince differ considerably from those just described for the interior, mainly with regard to orientation.

Along the northern boundary of the subprovince, within the supracrustal rocks north of the Wabigoon fault at Dryden (*see* Figure 9.8), asymmetric folding of stratigraphic units is the dominant structural style, with Z-styles

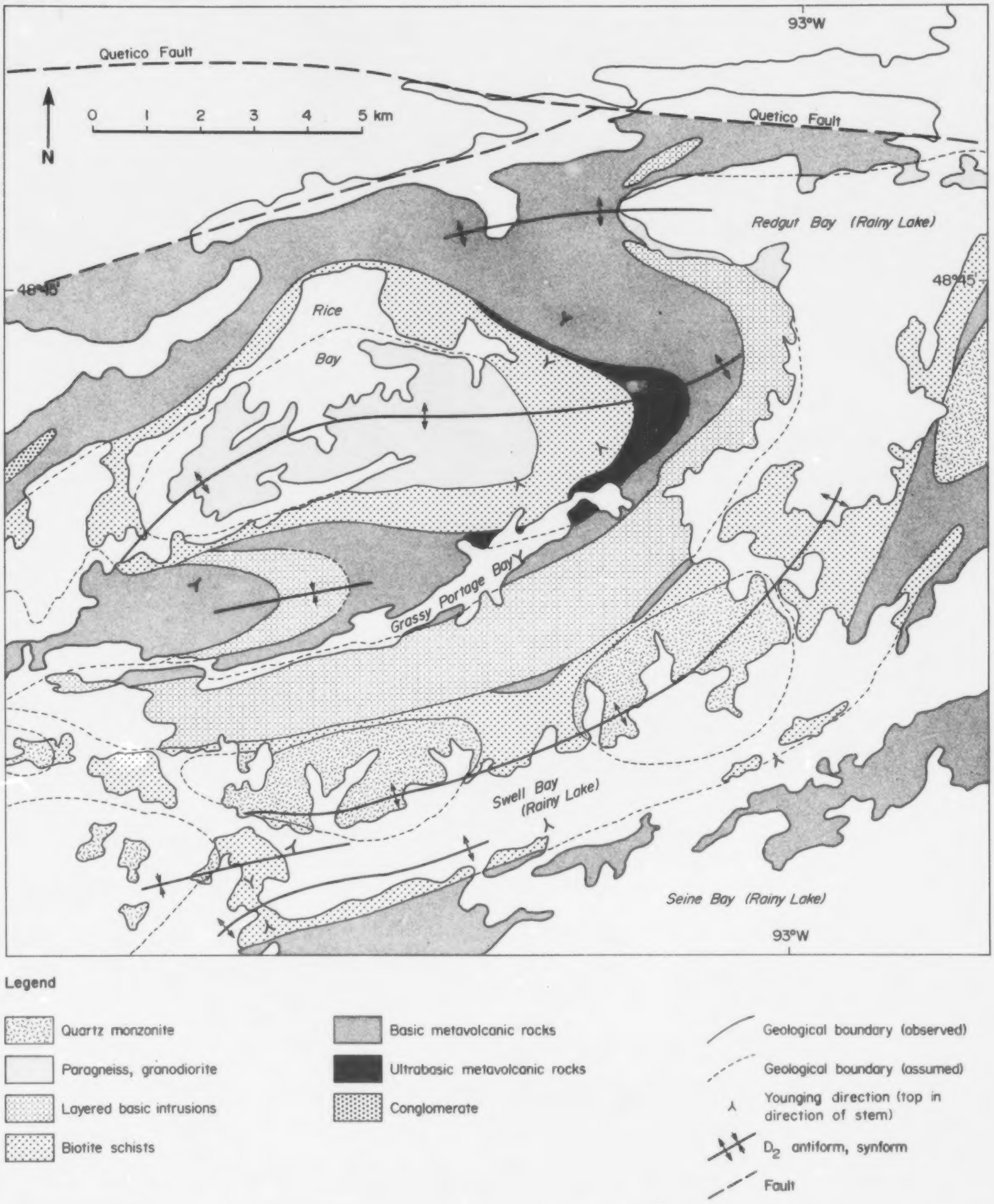

Quetico Fault
93°W
0 1 2 3 4 5 km
N
Quetico Fault
Redgut Bay (Rainy Lake)
48°45'
48°45'
Rice Bay
Grassy Portage Bay
Swell Bay (Rainy Lake)
Seine Bay (Rainy Lake)
93°W
Legend
Quartz monzonite
Paragneiss, granodiorite
Layered basic intrusions
Biotite schists
Basic metavolcanic rocks
Ultrabasic metavolcanic rocks
Conglomerate
Geological boundary (observed)
Geological boundary (assumed)
Younging direction (top in direction of stem)
D2 antiform, synform
Fault

Figure 9.36d. Examples of structures in the Wabigoon Subprovince. Geological map showing the relationship between younging directions and D_2 folds in the Rice Bay area, Rainy Lake (*modified from* Poulsen et al. 1980.

dominating over S-styles. A first generation of minor, tight, steeply west-plunging Z-folds occur adjacent to the fault; pegmatite and quartz veins are offset dextrally; a second generation of minor, open, steeply southwest-plunging Z-folds refolds the earlier set. The shape and orientation of the second generation Z folds are similar to major fold structures north of the Wabigoon fault at Thunder Lake. This regional-scale fold asymmetry corresponds spatially with the northern margin of the subprovince and may be interpreted as being due to a component of dextral strike-slip

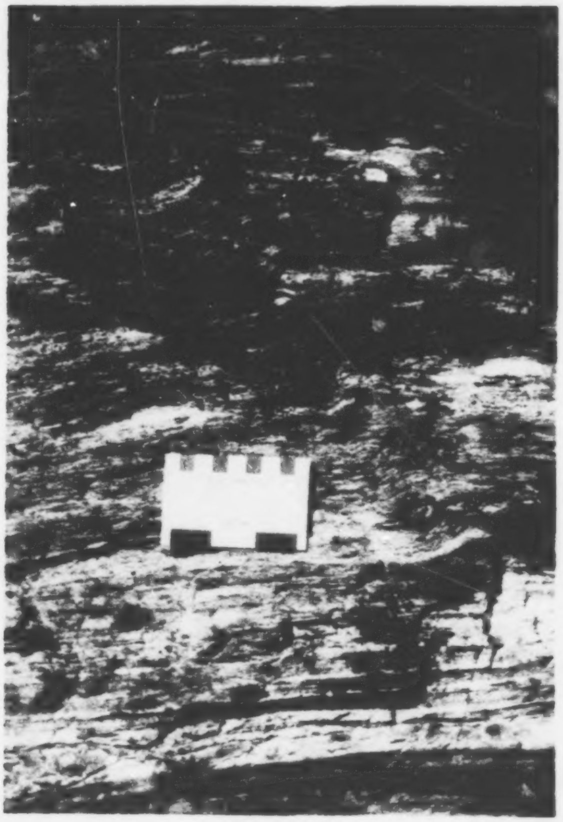


Figure 9.37. Late brittle deformation (kink bands) along the Wabigoon fault, Piskegomang Brook (*see* Figure 9.10 for location; photograph *courtesy of* C. Blackburn).

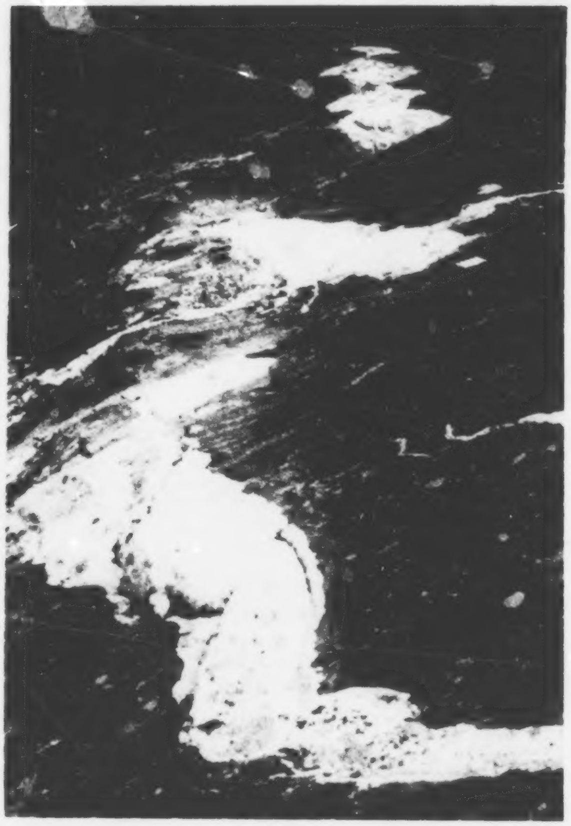


Figure 9.39. Dextral offset of a pegmatite vein in the Warclub group metasedimentary rocks, along the Wabigoon fault, in a municipal park in Dryden (photograph *courtesy of* G. Johns).

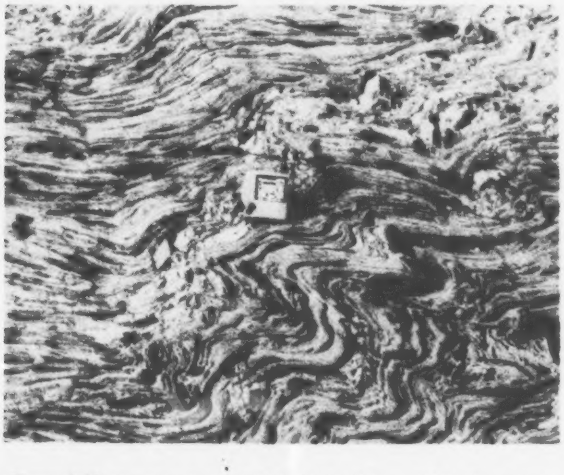


Figure 9.38. Complex ductile folding in the Monte–Cristo shear zone through the lower member of the Cameron Lake volcanics, near Rowan Lake (photograph *courtesy of* G. Johns).

shear marginal to the Wabigoon fault. The fault marks the southern boundary of this deformation style (Blackburn 1979; Blackburn et al. 1982); no examples of these folds have been recognized south of the fault, where folds have subhorizontal axes. Dextral movement is therefore considered to have been accommodated by folding in supracrustal units north of, and accompanied by strike-slip movement along, the fault.

In the Sandybeach Lake area (*see* Figure 9.12) between Dryden and Sioux Lookout (*see* Figure 9.2), Chorlton (1991) documented 4 discrete stages of deformation (*see* Table 9.11). Detailed mapping of iron formation units and the contact between the Minnitaki and Neepawa groups allowed Chorlton to define flat-lying folds (D_1) and an associated, premetamorphic axial planar schistosity. The schistosity was subparallel with or slightly inclined to lithologic layering and major stratigraphic contacts. A second stage of deformation (D_2) is temporally and spatially associated with pluton emplacement, producing contact strain and thermal aureoles. Plutons are surrounded by a locally developed steeply dipping foliation and an aureole of amphibolite-facies metamorphism. Subsequently, a D_3 deformation stage of regional folding, along northeasterly trending axes, produced open to isoclinal upright folds that are doubly plunging. These have interfered with D_1 folds to produce complex outcrop patterns of domes and basins. The D_3 shear zones are steeply dipping, northeast trending and contain a southwest-plunging mineral lineation that, in addition to conjugate fracture systems, indicate oblique southeast-side-up, sinistral kinematics. A final stage of deformation (D_4) was continued convergence that produced the Wabigoon shear zone and associated splays, induced Z-style minor and mesoscopic folds of earlier fabrics and lithologic contacts, and reactivated dip-slip shears. Clockwise rotation of lithologic contacts, earlier fabrics, and shear zones indicate that the latest motion on the Wabigoon shear zone was with a dextral sense.

This structural development is in accord with those in other areas within the interior of the Wabigoon Subprovince and serves to indicate an early phase of recumbent folding, perhaps synchronous with premetamorphic tectonic interleaving of lithologic units of contrasting age and derivation. This was followed by progressive folding and shearing of this tectonic collage to form belt-parallel folds that postdated much of the magmatism (Chorlton 1991).

Farther to the west, in the Lake of the Woods region (*see* Table 9.11), Sanborn-Barrie (1991b) identified 2 phases of deformation (D_1 and D_2) north of the Crowduck Lake–Witch Bay deformation zone adjacent to the Winnipeg River Subprovince (*see* Figure 9.9). South of this deformation zone, Ayer and Buck (1989) also recognized the above 2 phases of deformation in a region well removed from the boundary (*see* Table 9.11). Both D_1 and D_2 appear to be a response to north-directed compression and Sanborn-Barrie (1991b) considered that D_2 developed simultaneously in both the WWR and the Winnipeg River Subprovince.

Within the extreme western part of the Wabigoon Subprovince the unusually broad zone, containing east-trending rocks, folds and shear zones resulting from subprovincial collision, may in part be explained by the proximity there of the Wabigoon–Winnipeg River subprovince boundary and the Crowduck Lake–Witch Bay deformation zone. These 2 structures created local conditions that promoted the development of a tectonic style typical of subprovincial interaction. In other areas of the WWR, subprovincial interaction is typically confined to a narrower zone adjacent to subprovince boundaries.

Between Lac des Mille Lacs and Rainy Lake (*see* Figure 9.2), the Quetico Fault cuts through and forms the southern boundary of the Wabigoon Subprovince. Along it are many occurrences of Z-style asymmetric folds, that dominate over S-style. In addition, deflection of lithologic units into the fault zone led Mackasey et al. (1974) to suggest a major component of dextral strike-slip movement along this boundary zone. In the Rainy Lake region (Poulsen 1984b; Fumerton 1985), the Seine River fault splays southwestwards from the Quetico Fault; motion along the Quetico Fault is transferred to the Seine River fault which forms the original Wabigoon–Quetico subprovince boundary.

Similar evidence of initial dip-slip motion, superseded by dextral strike-slip movement is seen in the Beardmore–Geraldton Belt, where iron formation within wackes display complex but overall Z-style asymmetry in an otherwise east-trending, linear zone of alternating thrust-repeated mafic volcanic and clastic metasedimentary units (Devaney and Williams 1989; *see* Williams, this volume).

Fault Development at Subprovince Margins

Major transcurrent faults occur at and adjacent to the northern and southern edges of the subprovince, paralleling the structural grain that commonly parallels the subprovince margin (e.g., Paint Lake and Blackwater River faults, Devaney and Williams 1989). These faults are not always developed as wide zones of shearing but are commonly subtle, almost cryptic, and may be inferred from the juxtaposition of rocks of dissimilar age or type. An example is the Wabigoon fault, that occurs along the north side of the subprovince. It may be traced a distance of 250 km from the vicinity of Sioux Narrows, in the west (*see* Figure 9.10), to south of Minnitaki Lake, in the east (*see* Figure 9.12) and along its entire length coincides with the boundary between metasedimentary rocks of the Warclub and Minnitaki groups to the north and volcanic rocks to the south. Other examples along the north side of the subprovince are the Little Vermilion fault (*see* Figure 9.12), which passes into the Miniss River Fault to the northeast (*see* Figure 9.3), and the Kashaweogama fault in the Savant Lake belt (*see* Figure 9.3) (Sanborn-Barrie 1989).

QUETICO FAULT

The Quetico Fault, over 200 km long, parallels and in part defines the southern boundary of the subprovince (Kennedy 1984). Along it, dextral transcurrent offset has been the major movement, estimated by various workers to be up to 128 km (Mackasey et al. 1974; Bau 1979; Kennedy 1984; *see* Williams, this volume). These estimates have been based on, for example, the correlation of an ultramafic pyroclastic unit north of the Quetico Fault at Atikokan, the Ashrock Formation, with a similar ultramafic pyroclastic unit, at Rainy Lake, shown to be almost identical in composition (Schaeffer and Morton , in press). The Quetico Fault differs from transcurrent faults within the Wabigoon Subprovince in a number of respects: 1) the fault does not follow lithologic boundaries but cuts across both supracrustal rocks, and major batholithic domes (e.g., Rainy Lake batholithic complex); 2) the fault is straight and is a complex zone of deformation; and 3) the fault contains rocks with a mylonitic fabric (Kennedy 1984). The fault is a major crustal feature and was operative over a protracted time period, into the Paleoproterozoic (Peterman and Day 1989).

THRUST FAULTS

Thrust faults, including high-angle reverse faults, have been recognized or inferred by a number of workers in different parts of the subprovince, predominantly toward the subprovince margins, but in at least 1 case, within the internal part of the subprovince. The cryptic nature of thrust faults is such that as yet unrecognized thrusts may be present within the interior of the subprovince.

Internal Thrust Faults

The interpretation of internal thrust faults is contingent on equivocal geochronologic evidence. The Mosher Bay–Washeibemaga Lake fault, in the Manitou Lakes greenstone belt, has been interpreted as a thrust. Boyer Lake mafic volcanic rocks lie with fault contact upon the pyroclastic to sedimentary Manitou and correlative Stormy Lake groups (Blackburn 1980, 1982). A felsic tuff towards the top of the Boyer Lake group has been dated at 2728 Ma, and a gabbroic intrusion into the Boyer Lake group at 2727 Ma, whereas felsic volcanic rocks within the structurally underlying Manitou and Stormy Lake groups have been dated at 2699 and 2703 Ma respectively.

An additional possible thrust fault within the interior of the subprovince is the Barrier Island fault in the Lake of the Woods greenstone belt (*see* Figure 9.9). There, north-facing mafic volcanic rocks of the Barrier Islands group are in fault contact with younger, north-facing pyroclastic rocks of the Wiley Bay group (Ayer 1989).

Marginal Thrust Faults

Evidence of thrust movement at the southern and northern margins of the subprovince is interpreted from geochronology; from sheared contacts between repetitive and intercalated supracrustal units; and from the occurrence of inverted stratigraphic units requiring recumbent fold and, or, nappe formation.

Thrust stacking in the Sioux Lookout greenstone belt is interpreted from geochronology; Neepawa group volcanic rocks (2733±1 Ma, Davis and Trowell 1982) structurally overlie sedimentary rocks of the Abram group that are intruded by a porphyry stock (2703±2 Ma, Davis and Trowell 1982). The Little Vermilion fault, separating these 2 groups, may therefore have both a transcurrent and a thrust component.

Tight folds with subhorizontal axes lying parallel to the Wabigoon fault, and at the top of the Upper Wabigoon volcanics at Eagle and Wabigoon Lake, are indicative of early thrusting of the Upper Wabigoon volcanic rocks over the metasedimentary rocks of the Warclub and Minnitaki groups (cf., Figure 4 in Blackburn 1980; Blackburn 1979; Blackburn et al. 1982).

The Wabigoon–Quetico subprovince boundary in the Beardmore–Geraldton area has been interpreted as a zone of south-verging thrusts (Williams 1986; Devaney and Williams 1989). The east-trending belts of alternating sedimentary and mafic volcanic rocks are interpreted as thrust-bound slices on the basis of shearing and isoclinal folding localized at lithologic contacts and sporadic kinematic indicators. On a regional scale, relatively proximal sedimentary facies have been juxtaposed structurally above more distal ones. Rotation of the thrust-bounded units to a vertical orientation was followed by inhomogeneous ductile shear with dextral sense kinematics, especially within pelitic and iron formation units and pre-existing thrust zones (Devaney and Williams 1989).

Inverted rocks within the Rice Bay Dome (*see* Figure 9.13) form part of a wedge lying between the Quetico and Seine River transcurrent faults (Poulsen et al. 1980; Davis et al. 1989). Structural analysis shows that mafic volcanic rocks at the edge of the dome face inward and, therefore, downward despite the fact that they structurally overlie metasedimentary rocks in the central part of the dome (*see* Figure 9.13). A long-debated controversy concerning the original relationships between these mafic metavolcanic rocks (Keewatin) and the metasedimentary rocks (Seine and Coutchiching) dates back to the first regional geological work in this area (Lawson 1913). The significance of this inverted, now domal, sequence is that it may represent either the lower inverted limb of a regional recumbent fold, or a thrust nappe of inverted strata (Poulsen et al. 1980; Davis et al. 1989).

Just within the Winnipeg River Subprovince at the northern edge of the Wabigoon Subprovince, a large-scale recumbent fold near Silver Lake may be further evidence of a large-scale, subhorizontally directed tectonic movement involving the development of nappes (Beakhouse et al. 1983).

GEOPHYSICAL EXPRESSION OF MAJOR STRUCTURES

In the Wabigoon Subprovince, in common with the other Archean granite-greenstone subprovinces, rock types with high magnetic susceptibility (iron formation and mafic and/or ultramafic intrusions) are clearly distinguishable on the aeromagnetic maps accompanying this volume. Major observations not reported elsewhere in this chapter are as follows:

1. The 30 km diameter circular array of mafic and/or ultramafic intrusions which includes the Lac des Iles intrusive complex (Sutcliffe 1986) is apparent.
2. There appears to be a greenstone septum linking the Sturgeon Lake and the Obonga Lake greenstone belts spanning the CWR.
3. The structural grain of the Beardmore–Geraldton area (EWR) continues beneath Lake Nipigon to the Lac des Iles area (Pye 1968).
4. The Quetico Fault crosscuts the trend of the inferred extension of east-trending thrusts of the Beardmore–Geraldton Belt (Williams 1990, this volume).
5. Major shear zones within the WWR, such as the Manitou fault, are visible.
6. Relatively continuous short wavelength magnetic highs and lows suggest many of the greenstone remnants within the CWR may be connected.
7. There is a consistent difference in magnetic susceptibility between internal batholiths and those external to the greenstone belts. Those within greenstone belts tend to exhibit relatively low susceptibilities (e.g., Atikwa–Lawrence and Aulneau), whereas external batholiths exhibit slightly higher magnetic susceptibilities. Those with higher susceptibilities tend to be granitic batholiths, whereas the internal batholiths, dominated by rocks of the foliated tonalite suite, are of lower magnetic susceptibility.

The 1:5 000 000 scale gravity map (CGAMNA 1988) shows the entire subprovince to have a relatively high gravity field, with a low in the area of the White Otter Lake batholith. The continuity of the subprovince as a major structure from the Geraldton area eastward to the Kapuskasing Structural Zone is evident on the map.

Gravity modelling shows the depth extent of the Sturgeon Lake greenstone belt to be generally about 3 km (Dusanowskyj 1976), similar to that reported for many other Superior Province greenstone belts (cf., Gupta et al. 1982). The depth extent of granitic batholiths is variable:

1. The Aulneau Batholith, has a depth extent of 5 to 7 km with relatively flat base and a central deeper part extending to 10 to 11 km (Hall and Brisbin 1982).
2. Batholiths in the Sturgeon Lake area (Dusanowskyj 1976) have a depth extent of about 16 km, whereas posttectonic syenite and granitic stocks have a depth extent of 2 to 5 km.
3. Gravity modelling in an area just south of Sturgeon Lake (Szewczyk and West 1976) shows the potassium feldspar megacrystic Indian Lake batholith to be a sheet-like body only 2 km deep, whereas the compositionally similar Basket Lake batholith has an 8 km depth extent.

The relationship between limited vertical extent and abundant potassium feldspar phenocrysts in granitic batholiths suggest that it is the deep interior part of late potassium-rich intrusions which contain potassium feldspar phenocrysts. From this relationship between depth extent and phenocryst abundance and the flat-bottomed morphology of the Aulneau and Sturgeon Lake area batholiths (1 and 2 above), we suggest that granitic batholiths are in part sheet-like in form with a relatively limited depth extent.

TIME STRATIGRAPHIC CORRELATION WITHIN THE WESTERN WABIGOON

Parts of the WWR have been the subject of attempts at characterizing "type" Archean stratigraphies (Lawson 1913; Goodwin 1970; Wilson and Morrice 1977; Blackburn et al. 1985). These attempts were based on recognizing consistency in rock type, superpositional relationships and deformation state. However, the structural complexity within Archean supracrustal sequences makes stratigraphic correlation within and between greenstone belts speculative. Modern U-Pb dating techniques, however, have precisions of 1 to 5 million years, much shorter than the time span for volcanism within many greenstone belts (less than 50 million years). The possibility of correlating widely separated units is now within reach. This section attempts to correlate supracrustal units within the Kakagi Lake–Savant Lake greenstone belt of the WWR, the only area where the degree and extent of mapping and geochronology might make this a realistic exercise. Compounding the problem of areal distribution of age determinations is their concentration within felsic units of middle and upper diverse sequences. However, improvements in analytical sensitivity permit precise age determinations, even on single grains, expanding the range of suitable samples and also permitting age determination on detrital grains in sedimentary rocks, which can give important age constraints and information on provenance (Davis et al. 1989; Davis 1990).

The approach used here is to construct structural sections in well-studied areas of the WWR, and then correlate units and sequences on the basis of absolute age constraints and consistent relationships with adjacent units. Lithologic similarity may be used but with caution, since few rock types are distinctive enough to be uniquely characteristic of a given time-equivalent sequence. The possibility of thrust stacking means that relative age relationships cannot be assured even within a single supracrustal sequence, unless contacts between adjacent units can be shown to have no significant displacement. Unfortunately, bedding- parallel faults are common in Archean supracrustal sequences and contacts are often poorly exposed. Therefore, the principle of stratigraphic superposition should also be used with extreme caution (cf., deWit et al. 1987).

The structural sections are illustrated in Figure 9.40, and Table 9.7 gives a detailed presentation of possible time-equivalent units and the characteristics defining them. The fundamental subdivision (*see* Table 9.7) of the supracrustal sequences is between those predating major deformation (type 1) and those which are synchronous or postdate it (type 2). The preorogenic supracrustal rocks are dominantly volcanic and include most of the lower and upper mafic sequences, and the upper diverse volcanic sequences. The synorogenic rocks are dominantly sedimentary and appear to include most of what were previously called Keewatin- and Timiskaming-type metasedimentary rocks (*see* Discussion *in* Metasedimentary Rocks). Relatively minor calc-alkalic and alkalic volcanic units are intercalated with the synorogenic metasedimentary rocks. Preorogenic plutonic rocks include the synvolcanic rim phases of internal batholiths and the large mafic intrusions. Synorogenic plutonic rocks include the potassium-rich syntectonic to late tectonic plutons and quartz-feldspar porphyry bodies spatially associated with the metasedimentary rocks. Much of the Aulneau and Dryberry batholiths may also be syntectonic, as well as some of the core phases of large internal batholiths such as the Atikwa–Lawrence batholith.

Type 1 preregional deformation rocks have ages from 2775 to 2718 Ma in the Kakagi Lake–Savant Lake greenstone belt. In contrast, zircon inheritance in type 2 syntectonic or postdeformation volcanic rocks commonly results in a complicated age pattern; age constraints can be established, as with metasedimentary rocks, from the youngest concordant zircon grains and from ages of cross-cutting intrusions. Based on these data, type 2 rocks show a relatively restricted time span and range in age from less than 2714 to greater than 2696 Ma.

Type 1 sequences have been divided into temporal subgroups, designated by letters in Table 9.7. The time intervals chosen for these subdivisions may reflect time-equivalent packages of rocks which evolved as part of the same; igneous complex during a discrete part of greenstone belt development. For example, units having different ages within the time division 1b (2745 to 2732 Ma) in some cases are contained within the same assemblage (e.g., Wabigoon volcanics) and are commonly structurally overlain by turbiditic metasedimentary rocks. Sequences in division 1d, although consisting of different rock types (basaltic flows—Boyer Lake group versus felsic pyroclastic rocks—Kakagi Lake group), all contain remarkably thick, differentiated mafic sills of similar age.

Data indicate that the lower mafic sequences are not all of the same age but form at least 2 groups: Type 1a, the older sequence, underlies type 1b rocks and a single dated example is 2775 Ma; and type 1b, a younger mafic sequence (e.g., the Katimiagamak group which appears laterally continuous with the Wapageisi group). Several of these sequences, including the Wapageisi, Jutten, and Botham Bay groups are overlain unconformably by type 2 alluvial metasedimentary rocks. Age data indicate that these unconformities are indistinguishable in age (2699 to 2696 Ma). The metasedimentary rocks are all structurally overlain by older type 1b and 1d sequences, presumably emplaced by thrusting. Furthermore, the fact that the submarine mafic sequences are overlain by alluvial metasedimentary rocks indicates that the mafic rocks may also have been tectonically transported into a subaerial environment, presumably along thrust faults. This, along with the similar structural position of these sequences at the margins of the greenstone belt and facing inward toward its axis, is a basis for correlating both lower mafic sequences and metasedimentary rocks, respectively, among several areas around the greenstone belt.

Type 2 rocks can be subdivided on lithologic grounds, but with 1 exception, there is not a consistent time difference among sequences in different parts of the belt. The exception is the Seine group, on the southern margin of the WWR, which was deposited in the age range 2693 to 2685 Ma and which postdates sedimentary sequences in the interior of the belt, many of which were intruded by late plutons at 2696 Ma.

Attempts at time-stratigraphic correlation may indicate whether greenstone belts consist of a collage of disparate and unrelated elements which are tectonically juxtaposed, or whether most of the sequences in the belt are autochthonous or spatially related and evolved as part of a single magmatic system. In the latter case, knowing which units are time correlative may help to determine chronologic evolution of the belt and enable a spatial and tectonic reconstruction of environments prior to deformation.

Evidence that most of the supracrustal sequences in the WWR formed as part of a single tectono-magmatic complex, which evolved over at least an 80 million year time span, is based on apparent stratigraphic conformity between type 1a and 1b sequences in the Sturgeon–Savant lakes and Wabigoon Lake areas, and between type 1b, 1c, and 1d sequences in the Kakagi Lake area. Type 1b sequence in the Kakagi Lake area is also intruded by a major batholith which is coeval with 1d (*see* Figure 9.40). These sequence types include most of the supracrustal units in the belt. In addition, although stratigraphic relations have been modified by regional deformation, time correlative units can be recognized in widely different parts of the belt and there is a remarkable degree of stratigraphic and structural consistency within and among different units throughout the belt.

Type 1a sequences may represent oceanic crust which perhaps formed in an older back-arc basin and on which the main 2745 to 2718 million-year-old arc edifice was deposited. The lower mafic 1b sequences may represent a back-arc basin contemporaneous with the 2745 to 2718 million-year-old arc. Collision with, and overthrusting of, type 1 rocks onto an older sialic block, presently exposed in the Winnipeg River Subprovince and the CWR, may have caused uplift and deposition of the type 2 metasedimentary rocks. This event forms part of an extended period of orogeny which appears to have occurred diachronously from north to south across the Superior Province (*see* Corfu and Davis, this volume).

Central Wabigoon Region Summary

The central region of the Wabigoon subprovince has not been lithostratigraphically subdivided in the same way as the WWR, as geochronologic data are scarce. The fundamental architecture of the region, involving dominantly gneissic and granitic rocks with subordinate septa of greenstones, was reviewed in an earlier section; here we offer some interpretation of the supracrustal rocks displayed on the Bedrock Geology of Ontario (*see* Maps 2542 and 2543, map case) and Tectonic Assemblages of Ontario (*see* maps 2576 and 2577, map case) maps accompanying the volume. The greenstones have not been classified with respect to depositional setting everywhere.

The Garden Lake and Bo Lake–Heaven Lake greenstone belts (*see* Figure 9.2; Thurston et al. 1987) have few felsic metavolcanic rocks, some komatiitic flows and are thus similar to the platformal greenstones of the Atikokan and Lumby Lake area (Thurston and Chivers 1990).

The Obonga Lake greenstone belt (*see* Figure 9.2) may be a composite greenstone belt composed of assemblages representing various depositional environments and possibly various ages as described below:

1. The southern part of the belt has rare felsic metavolcanic units and unconformably overlies (Cortis et al. 1988) tonalitic rocks cut by abundant amphibolite dikes; by analogy with the Atikokan area, these supracrustal rocks may be older than 2.8 Ga.
2. The central part of the Obonga belt represents a lower mafic sequence and an upper diverse sequence approximately 2730 million years in age.
3. A metasedimentary sequence in the north part of the belt is distinct from 1 and 2 above. It is composed of conglomerate with clasts of granitoid material, wacke, arkose, slate, phyllite and quartzite (Kustra 1966). This sequence may represent either a platformal or a Timiskaming-type (cf., Thurston and Chivers 1990) setting.
4. The northern margin of the belt is composed of the tectonically bounded, serpentinized Chrome Lake –Puddy Lake ultramafic body (Whittaker 1980).

The Caribou Lake greenstone belt in the CWR includes tholeiite and komatiite with subordinate felsic metavolcanic rocks (Sutcliffe 1988). This greenstone belt is bordered by 3 billion-year-old granitic rocks (Davis et al. 1988). The scarcity of felsic metavolcanic rocks, the presence of komatiites and the age of the belt suggest comparison with the Steep Rock Lake area.

The CWR consists of some units about 3 billion years old; stratigraphic relationships indicate that part of the region is a basement complex. However, the extent of the basement complex is uncertain and the presence of 2.7 billion-year-old units suggests that there is a long complex history to be unravelled in this region.

Eastern Wabigoon Region Summary

Supracrustal rocks of the eastern region of the Wabigoon Subprovince have been subdivided into mafic and diverse sequences which are described above. However, some regionally significant dissimilarities between the 2 regions are apparent. In Barbara, Meader, Pifher (Kresz and Zayachivsky 1989) and Lapierre townships (Kresz 1989) an unusual subaerial, south-younging andesite to rhyolite sequence overlies basal tholeiites. The sequence predates movement along the Paint Lake deformation zone (Williams 1991, this volume) separating the Beardmore–Geraldton Belt from the Onaman–Tashota terrane; therefore, the sequence is probably not of the Timiskaming type. It may represent a diverse sequence similar to the WWR; however, it resembles, in terms of rock types, environment of deposition and relationship to tectonic events, the late andesite-bearing sequences of the Australian shield (Hallberg et al. 1976) which are a type of arc sequence.

The metasedimentary rocks near Conglomerate Lake (Amukun 1980) include sandstones and conglomerate with abundant granitic clasts and marked lateral variation in clast type. The clast types and variation and the position of the unit near a major change in the strike of the surrounding metavolcanic rocks suggests that the metasedimentary unit may be comparable to Timiskaming-type sequences of the WWR. The O'Sullivan Lake area in the northern part of the region lacks major felsic metavolcanic units and includes a high proportion of tholeiites (Inasi 1981). The lack of felsic metavolcanic units and metasedimentary rocks suggests that the greenstones of the O'Sullivan Lake area do not represent an upper diverse sequence. They may be similar to the greenstones of the Caribou Lake area (Sutcliffe 1988) or represent a lower mafic sequence such as those in the WWR. In the Onaman Lake area, 2764^{+6}_{-5} million-year-old (Anglin et al. 1988) felsic volcanic rocks represent a shallow-water tuff-cone environment (Osterberg et al. 1987).

The aforementioned unusual sequences are but a small part of the EWR but illustrate that the EWR may be interpreted in terms of mafic sequences and diverse sequences upon which Timiskaming and andesitic units were deposited. The EWR is distinguished from the WWR by unusual sequence types and possibly by a greater age of volcanism, based on limited data.

METAMORPHISM

Introduction

In the Wabigoon Subprovince, there is little evidence of high metamorphic grades within the supracrustal and granitic rocks. Although pelitic rocks are relatively uncommon, only low-grade metamorphic mineral assemblages have been found in them. In volcanic rocks, middle amphibolite metamorphic assemblages are the highest recorded (Ayres, 1978). Distribution of metamorphic grades is spatially associated with batholiths; highest metamorphic grades are adjacent to these major structures.

Metamorphic grades and patterns contrast with those in the Quetico, Winnipeg River, and English River subprovinces, where low-pressure granulite mineral assemblages, migmatization and anatexis are common. Grades of metamorphism increase at the margins of Wabigoon Subprovince, into the adjoining subprovinces.

Conditions of Metamorphism

Pressures during metamorphic recrystallization have been estimated using characteristic mineral assemblages. Altered volcanic rocks in the interior of Wabigoon Subprovince contain kyanite-andalusite-, sillimanite-staurolite- (Lefebvre 1982) and kyanite-andalusite-chloritoid- (Franklin et al. 1975) bearing assemblages indicating pressures of 0.3 to 0.4 GPa. At the margins of the Wabigoon subprovince, assemblages containing cordierite, andalusite and sillimanite (Bartlett 1978), and staurolite, garnet, sillimanite and andalusite (Poulsen 1984a) indicate pressures of 0.3 to 0.4 GPa. In adjoining parts of the Quetico Subprovince, staurolite-andalusite-garnet-bearing assemblages (Pirie and Mackasey 1978) indicate pressures of 0.3 to 0.4 GPa.

Temperatures during metamorphism range up to about 750°C.

Timing of Metamorphism

Peak conditions of metamorphism in the subprovince boundary zones appear to have postdated major folding deformation; mappable isograds cross major fold structures (Bartlett 1978; Pirie and Mackasey 1978; Poulsen 1984a). Peak metamorphic grade was reached after inversion of the rocks within the Rice Bay Dome (Poulsen et al. 1980).

MINERAL DEPOSITS

Introduction

This section provides a brief review of the significant types of mineral deposits in the Wabigoon Subprovince, describes the spatial and temporal patterns of mineral deposit distribution and uses examples from throughout the subprovince. The major deposit types are listed in Table 9.12 and metallogenic trends in the Superior Province are described in Fyon et al., this volume. Deposits are described within the lithologic framework which is the basis for the organization of this chapter.

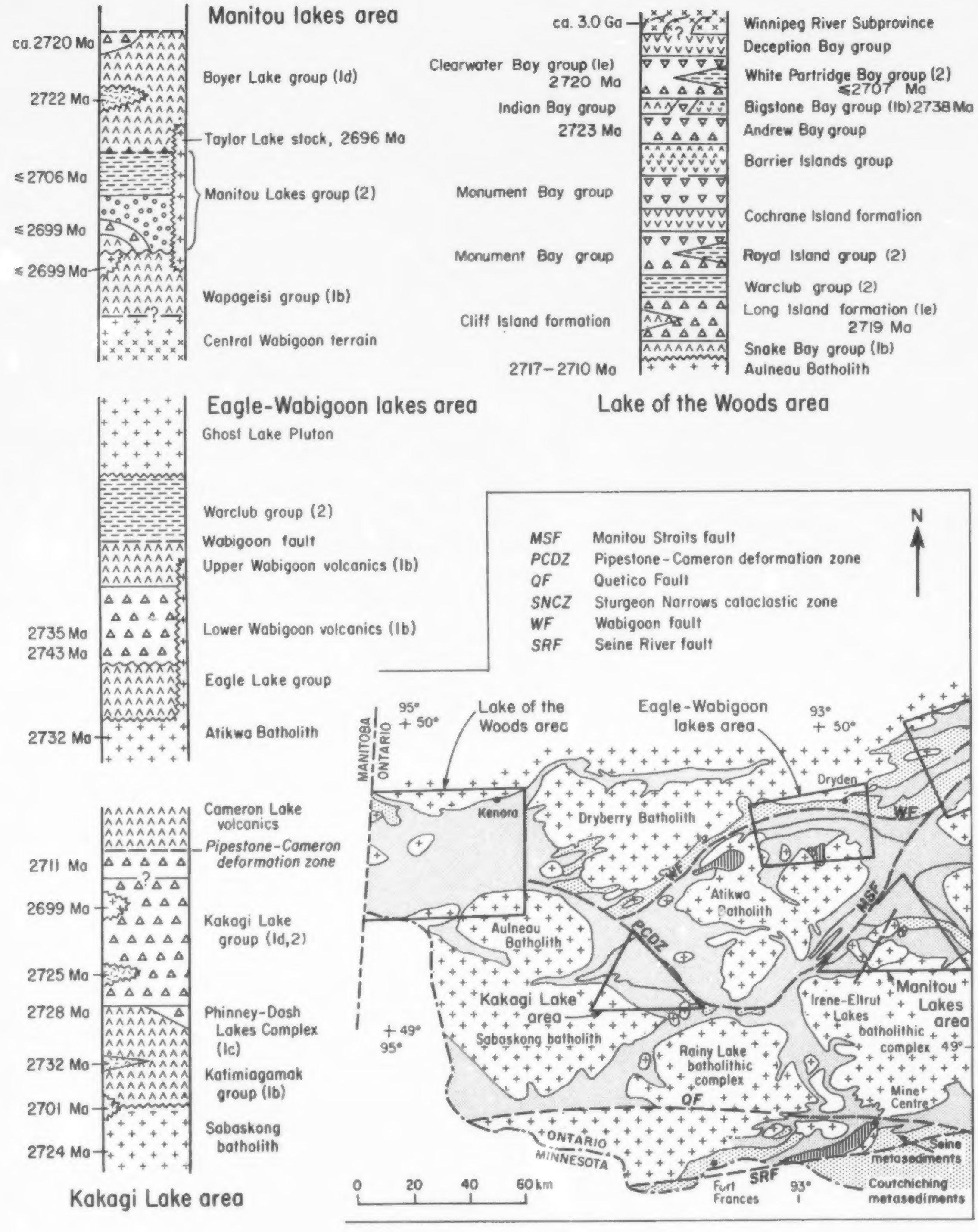

Manitou lakes area
ca. 2720 Ma
Boyer Lake group (Id)
2722 Ma
Taylor Lake stock, 2696 Ma
≤2706 Ma
Manitou Lakes group (2)
≤2699 Ma
≤2699 Ma
Wapageisi group (Ib)
Central Wabigoon terrain
ca. 3.0 Ga
Winnipeg River Subprovince
Deception Bay group
Clearwater Bay group (Ie) 2720 Ma
White Partridge Bay group (2) ≤2707 Ma
Indian Bay group 2723 Ma
Bigstone Bay group (Ib) 2738 Ma
Andrew Bay group
Barrier Islands group
Monument Bay group
Cochrane Island formation
Monument Bay group
Royal Island group (2)
Warclub group (2)
Cliff Island formation
Long Island formation (Ie) 2719 Ma
Snake Bay group (Ib)
2717–2710 Ma
Aulneau Batholith
Lake of the Woods area
Eagle-Wabigoon lakes area
Ghost Lake Pluton
Warclub group (2)
Wabigoon fault
Upper Wabigoon volcanics (Ib)
2735 Ma
2743 Ma
Lower Wabigoon volcanics (Ib)
Eagle Lake group
2732 Ma
Atikwa Batholith
Cameron Lake volcanics
Pipestone-Cameron deformation zone
2711 Ma
2699 Ma
Kakagi Lake group (Id,2)
2725 Ma
2728 Ma
Phinney-Dash Lakes Complex (Ic)
2732 Ma
Katimiagamak group (Ib)
2701 Ma
Sabaskong batholith
2724 Ma
Kakagi Lake area
MSF Manitou Straits fault
PCDZ Pipestone-Cameron deformation zone
QF Quetico Fault
SNCZ Sturgeon Narrows cataclastic zone
WF Wabigoon fault
SRF Seine River fault
N
95° + 50°
Lake of the Woods area
Eagle-Wabigoon lakes area
93° + 50°
MANITOBA
ONTARIO
Kenora
Dryberry Batholith
Dryden
WF
WF
Atikwa Batholith
MSF
Aulneau Batholith
PCDZ
Kakagi Lake area
Manitou Lakes area
Irene-Eltrut Lakes batholithic complex
49°
+ 49° 95°
Sabaskong batholith
Rainy Lake batholithic complex
Mine Centre
OF
ONTARIO
MINNESOTA
Seine metasediments
SRF
Fort Frances
93°
Coutchiching metasediments
0 20 40 60 km

Figure 9.40. Stratigraphic correlation of supracrustal sequences and plutonic units, Wabigoon Subprovince.

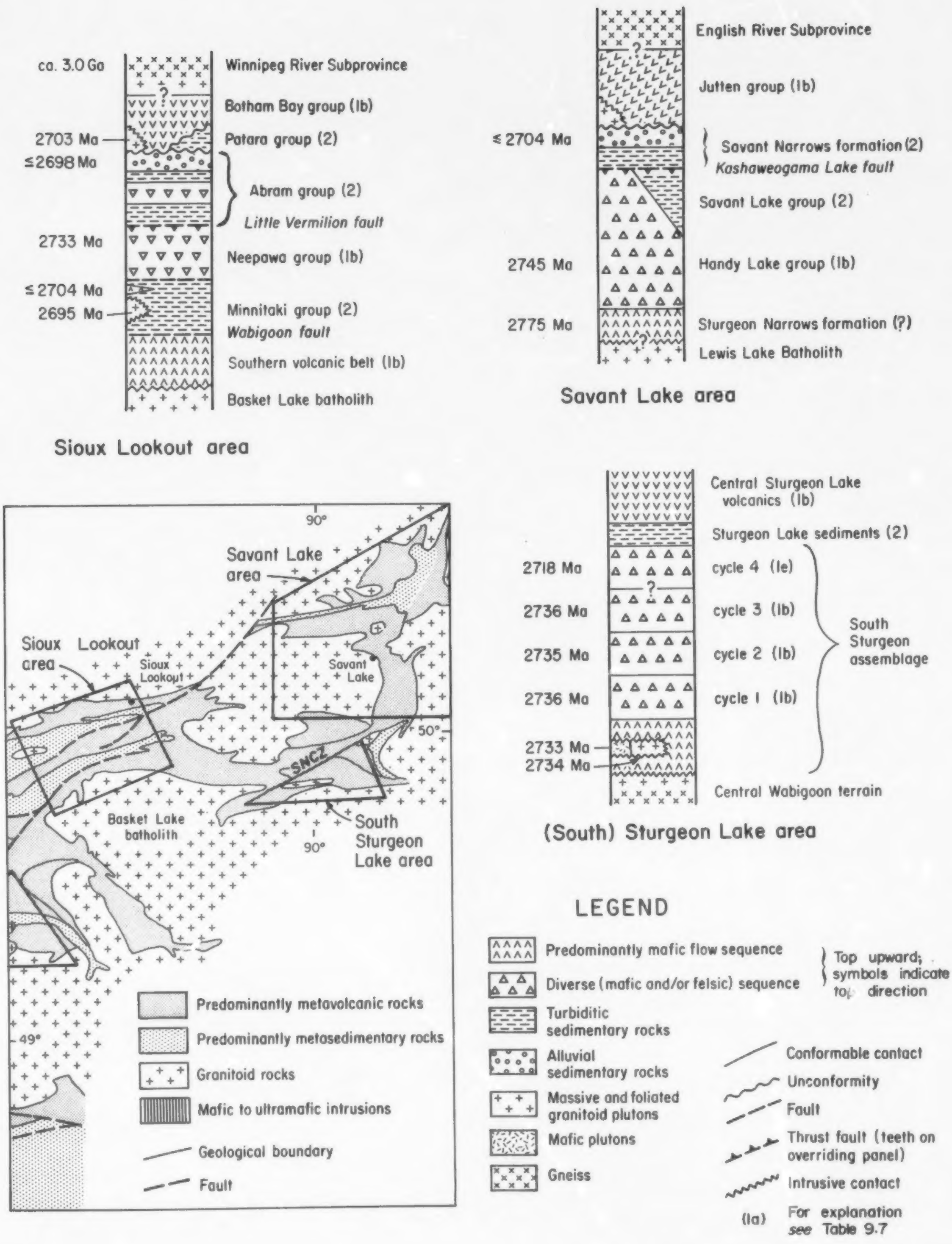

ca. 3.0 Ga
Winnipeg River Subprovince
Botham Bay group (Ib)
2703 Ma
Patara group (2)
≤2698 Ma
Abram group (2)
Little Vermilion fault
2733 Ma
Neepawa group (Ib)
≤2704 Ma
2695 Ma
Minnitaki group (2)
Wabigoon fault
Southern volcanic belt (Ib)
Basket Lake batholith
Sioux Lookout area
English River Subprovince
Jutten group (Ib)
≤2704 Ma
Savant Narrows formation (2)
Kashaweogama Lake fault
Savant Lake group (2)
2745 Ma
Handy Lake group (Ib)
2775 Ma
Sturgeon Narrows formation (?)
Lewis Lake Batholith
Savant Lake area
Central Sturgeon Lake volcanics (Ib)
Sturgeon Lake sediments (2)
2718 Ma
cycle 4 (Ie)
2736 Ma
cycle 3 (Ib)
2735 Ma
cycle 2 (Ib)
2736 Ma
cycle 1 (Ib)
South Sturgeon assemblage
2733 Ma
2734 Ma
Central Wabigoon terrain
(South) Sturgeon Lake area
90°
Savant Lake area
Sioux Lookout area
Sioux Lookout
Savant Lake
50°
SNCZ
Basket Lake batholith
90°
South Sturgeon Lake area
49°
Predominantly metavolcanic rocks
Predominantly metasedimentary rocks
Granitoid rocks
Mafic to ultramafic intrusions
Geological boundary
Fault
LEGEND
Predominantly mafic flow sequence
Diverse (mafic and/or felsic) sequence
Top upward; symbols indicate top direction
Turbiditic sedimentary rocks
Alluvial sedimentary rocks
Massive and foliated granitoid plutons
Mafic plutons
Gneiss
Conformable contact
Unconformity
Fault
Thrust fault (teeth on overriding panel)
Intrusive contact
(Ia) For explanation see Table 9.7

Figure 9.40. Stratigraphic correlation of supracrustal sequences and plutonic units, Wabigoon Subprovince.

Table 9.12. Metallogenic associations in Wabigoon Subprovince.

Syngenetic Deposits Supracrustal Sequence Type	**Commodities**	**Deposit Description**	**Deposit Type**	**References**
Syngenetic Deposits				
Metavolcanic Rocks				
Mafic	Cu-Au	Atikwa Lake massive to disseminated pyrrhotite (po), chalcopyrite (cp), minor pyrite (py), sphalerite (sp) and Au in interpillow space and fractures grading 1.12% Cu, 0.03 oz/ton Au	Cyprus-type volcanogenic massive sulphide (VMS)	Setterfield et al. 1983; Morrice 1977
	Cu-Zn	Rice Bay area sp, cp in sulphide-rich interflow argillites	Besshi-type VMS	Poulsen 1984a
Upper Diverse	Zn-Cu	Sturgeon Lake area sp, cp, po, py in 2735 million-year-old pyroclastic breccias as stratiform, strata-bound lenses in a caldera setting	Mattabi-type VMS	Trowell 1983; Morton and Franklin 1987; Morton et al. 1990
		Onaman Lake area sp, galena (gn), cp, Ag as stratiform, strata-bound lenses and crosscutting veins in hydroclastic 2764^{+6}_{-5} million-year-old felsic metavolcanic rocks	Noranda-type VMS	Osterberg et al. 1987; Anglin et al. 1988
		Marshall Lake area sp, cp, po, py in felsic pyroclastic rocks as deformed stratiform, strata-bound lenses	Noranda-type? VMS	Amukun 1989
Metasedimentary Rocks				
Platform metasedimentary rocks	Fe-Mn	Steep Rock Group Archean banded iron formation, supergene enriched during Archean and/or Cretaceous weathering; 527 000 000 tons	Algoma-type platformal iron formation	Machado 1987; Riley 1970; Shklanka 1972
Marine?		Bending Lake banded oxide iron formation in turbidite environment	Algoma-type deep marine iron formation	Shklanka 1968
Igneous Deposits				
Granitic Rocks				
Granite-granodiorite suite	U	Sioux Lookout, Vermilion Bay area, Rainy Lake weakly fractionated pegmatites associated with granite-granodiorite suite	uraniferous pegmatite	Blackburn 1973; Breaks 1982
Granite-granodiorite suite, foliated tonalites	Mo	Dryden area pegmatites related to the Revell batholith	pegmatite Mo	Kresz 1987
Granite-granodiorite suite	building stone	massive granite, minor potassium feldspar phenocrysts few joints		Blackburn et al. 1989
Two-mica granites	rare metals Li, Be, Cs, Nb, Ta, W	Dryden area strongly fractionated pegmatites in aureoles around S-type granite	rare metal (fractionated) pegmatite	Breaks 1989

Table 9.12. Metallogenic associations in Wabigoon Subprovince

Syngenetic Deposits Supracrustal Sequence Type	Commodities	Deposit Description	Deposit Type	References
Mafic-ultramafic Intrusions	Cu-Ni, platinum group elements, Au	Lac des Iles cp, po, pentlandite (pent), py, platinum group metals associated with late hydrothermal breccia along contact between gabbro and gabbronorite; 20.4 M tons of 6.34 g/t	pipe-type mineralization	Sutcliffe et al. 1989; Macdonald 1987
	Cu-Ni	Atikwa intrusive complex cp, pent, py, po associated with mafic-ultramafic stocks related to Atikwa–Lawrence batholith—noritic host rock	late breccia	Blackburn 1981; Davies 1973; Blackburn et al. 1989
	Cu-Ni	Rainy Lake area cp, pent at base of Grassy Portage intrusion; cp near top of Grassy Portage intrusion		Blackburn 1981; Poulsen 1984a
	Fe-Ti	magnetite (mag), Ti-mag near top of Grassy Portage intrusion and Bad Vermilion anorthosite		
	Cr	Chrome Lake disseminated, layered and podiform chromite near NE contact of intrusion	Alpine-type chromite mineralization	Whittaker 1986
Epigenetic Deposits	Au	Shoal Lake mineralization in shear zone transecting supracrustals in strain aureole of Snowshoe Bay batholith	contact strain aureole	Smith 1986
	Au	Cameron Lake quartz-carbonate breccia veins in shear zone in supracrustals	shear zones within greenstone belts	Melling et al., 1986
	Au	Sandybeach Lake area a) qtz ankerite vein in shear zones b) chlorite-ankerite schist and qtz veins c) qtz veins and late intrusions, minor ankerite	related to shear zones developed in late regional shortening within greenstone belt	Chorlton 1988
	Au	Beardmore–Geraldton and Mine Centre areas qtz–vein style mineralization in various supracrustal units, spatially associated with the Wabigoon–Quetico subprovince boundary	shear zones closely related to late shearing along subprovince boundary	Williams 1986, 1990; Poulsen 1984a
	Au	Beardmore–Geraldton area: pyritized iron formation localized near late shear zones	sulphidic replacement type	Macdonald 1988

Syngenetic Deposits

MAFIC SEQUENCES

The Atikwa Lake (*see* Figure 9.10) and Rice Bay area deposits (*see* Table 9.12 and Figure 9.9) represent volcanogenic sulphide deposits associated with mafic volcanism. In the former case, the hydrothermal system was operative on the scale of individual flows (Setterfield et al. 1983), and in the latter case, interflow metasedimentary rocks are mineralized. At Atikwa Lake, the alteration of the Maybrun deposit country rock basalts involves enrichment in iron, magnesium and potassium, and depletion in calcium and sodium in a fashion comparable to alteration patterns near volcanogenic massive sulphide deposits in Noranda (Gibson et al. 1983).

UPPER DIVERSE SEQUENCES

The major deposit type in the upper diverse volcanic sequences is the zinc-copper volcanogenic massive sulphide deposit (*see* Table 9.12). The most important deposits are in the Sturgeon Lake area associated with the 2735±1.2 million-year-old (Davis et al. 1985) South Sturgeon Lake volcanics (Trowell 1983). Six zinc-copper-silver-lead massive sulphide deposits make up the camp. These deposits are hosted by felsic, monolithic, clast-supported pyroclastic breccias. The deposits are all within a 30 by 4.5 km caldera (Morton et al. 1990) formed by: 1) initial caldera collapse in a subaerial to shallow-water environment; 2) subaqueous silicic explosive volcanism and ore development; and 3) terminal dome building with sedimentary and lava-flow fill and burial. Alteration zones are either widespread and conformable or are local lenses or pods beneath deposits and associated synvolcanic faults (Morton and Franklin 1987).

In the Onaman River area, vein and disseminated zinc-lead-copper-silver massive sulphides occur in the narrow belt of 2764^{+6}_{-5} million-year-old (Anglin et al. 1988) felsic metavolcanic rocks separating the Oboshkegan–Kowkash domain from the Elmhist–Castlewood domain. These occurrences are found at the top of 2 to 3 km by 200 m lenses of massive felsic lithic and crystal tuff. The deposits are related to tuff-cone volcanism in a shallow-marine to subaerial environment (Osterberg et al. 1987).

The Marshall Lake volcanic centre within the Marshall–O'Sullivan lakes terrane has massive sulphide mineralization. Amukun (1989) has described a stratabound horizon of copper-zinc-silver-lead occurrences in an intermediate to felsic tuff to lapilli tuff horizon (Figure 9.41 and 9.42) capped by ferruginous chert.

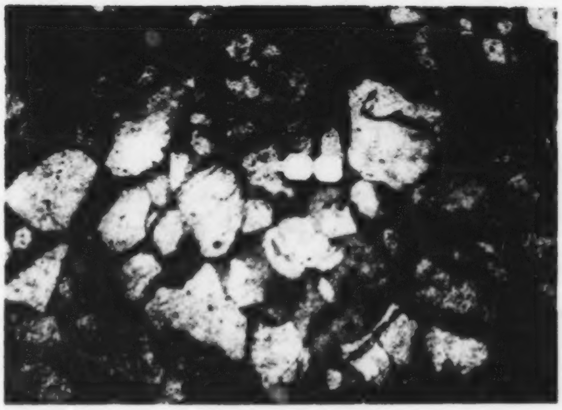


Figure 9.41. Dacitic pyroclastic breccia on the shore of Little Marshall Lake in the eastern Wabigoon region (photograph *courtesy of* S.E. Amukun).

METASEDIMENTARY DEPOSITS

Mineral deposits of sedimentary origin in the Wabigoon Subprovince are confined to Algoma-type (Gross 1965) chemically precipitated iron formation. Although magnetite iron formation layers within Keewatin-type clastic sedimentary rocks are abundant throughout the subprovince, paleoenvironmental work has been done only on the Savant Lake (Shegelski 1978) and Steep Rock Group (Wilks and Nisbett 1988). None have been of sufficient tonnage and grade to be commercially exploitable except the Steep Rock deposit. Goethite and hematite ores totalling over 700 Mt of reserves were identified in the Steep Rock Lake (Atikokan) area (Shklanka 1968).

The ore zone at Steep Rock Lake is part of the Joliffe Ore Zone Formation (Wilks and Nisbet 1988), a 100 to 400 m-thick unit consisting of the Manganiferous Paint Rock Member and the overlying Goethite Member (*see* Table 9.6). The Manganiferous Paint Rock contains "Buckshot Ore", which consists of hematite pisolites produced during Archean (Riley 1970) or Cretaceous (Machado 1987) weathering.

Origin of the Steep Rock Lake ores remains controversial; a secondary alteration origin was favoured by Bartley (1940), Gruner (1956), Moore (1938, 1940) and Shklanka (1972). A syngenetic origin was favoured by, among others, Joliffe (1955) and Wright (1965). A replacement origin was favoured by Bartley (1948), Hicks (1950), Roberts and Bartley (1943) and Tanton (1946). Recently, Wilks and Nisbet (1988) suggested the Mosher Carbonate was affected by development of karst topography (Joliffe 1955), so perhaps the Manganiferous Paint Rock Member is a paleosol. Weathering may be of Archean (Riley 1970) or Cretaceous (Machado 1987) age. A deposit occurs at Bending Lake (Figure 9.43).

Pluton-Related Deposits

INTRODUCTION

Granitoid-associated metallic mineral deposits of Wabigoon Subprovince are related to the massive granite to granodiorite suite and two-mica granites. Uraniferous pegmatites are related to both two-mica granites and to granite to granodiorite suite units (Breaks 1982). The most prolific area is marginal to the Winnipeg River Subprovince, between Kenora and Dryden. Uranium-bearing pegmatites are also present close to the southern boundary of Wabigoon

Subprovince, in the Rainy Lake area (Blackburn 1973). Molybdenite-bearing pegmatites are related to granite to granodiorite suite units (Kresz 1987). Molybdenite-bearing pegmatites are found at the margins of large internal batholiths, notably the Atikwa Batholith (Blackburn 1976b) and the Revell batholith (Kresz 1987). Molybdenite is also associated with the relatively sodium-rich synvolcanic stocks (e.g., Lateral Lake stock, Breaks et al. 1985).

Units of the massive granite to granodiorite suite are exploited for building stone near Dryden (Blackburn et al. 1989), where Nelson Granite Ltd., and Granite Quarriers (GQI) Inc. have successfully extracted and marketed a massive, pink, medium-grained, equigranular biotite granite.

Breaks et al. (1985) related the rare-metal-bearing pegmatites at the northern margin of the Wabigoon Subprovince to the two-mica granites developed within the Warclub–Minnitaki–Abram metasediments. Tungsten mineralization is also associated with the rare-metal pegmatites.

DEPOSITS IN MAFIC AND/OR ULTRAMAFIC INTRUSIONS

Mineralization associated with the mafic to ultramafic intrusions of the Wabigoon Subprovince comprises copper- and nickel-bearing sulphides, PGE, chromite and gold. As well as being scattered throughout the subprovince, there are 4 general areas of significant concentrations of these types of mineralization: 1) PGE mineralization within the Lac des Iles intrusive complex; 2) nickel-copper sulphides around the Atikwa Batholith; 3) the Rainy Lake area; and 4) chromite disseminations and seams.

Lac des Iles

The Lac des Iles mineralization is associated with a late igneous breccia up to 100 m wide between gabbro and gabbronorite (Sutcliffe et al. 1989). The mineralization totals 20.4 Mt with an average grade at 6.34 g/t PGE (Sutcliffe et al. 1989). The PGE mineralization at Lac des Iles is associated with disseminated chalcopyrite, pyrrhotite, pentlandite, pyrite and altered silicates (Sutcliffe et al. 1989). The breccia is cut by gabbro pegmatite dikes, locally containing up to 37 ppm total PGE and Au.

Sutcliffe et al. (1989) suggest that the mineralization was localized by mixing the PGE- and sulphide-rich gabbronorite and pyroxene cumulates with the volatile-rich gabbro. The PGE were redistributed by late volatile activity, which generated mineralized gabbroic pegmatites and breccia zones.

Atikwa Lake

A large number of nickel-copper sulphide occurrences occur within mafic to ultramafic intrusions surrounding the Atikwa Batholith (Mackasey et al. 1974). Information on occurrences west of the batholith is reported in Davies (1973a) and Blackburn et al. (1989). Partial remobilization of sulphide is apparent in a number of occurrences (Blackburn et al. 1989). The Kenbridge Mine, located west of the Atikwa Batholith, hosts the largest nickel-copper deposit within the Wabigoon Subprovince with reserves of 3.3 Mt grading 1.06% Ni and 0.54% Cu (Davies 1973a). Mineralization consists of pyrrhotite, pentlandite and chalcopyrite mainly hosted by a breccia matrix. Sulphides occur as net-textured ore, massive ore, blebs and fracture fillings.

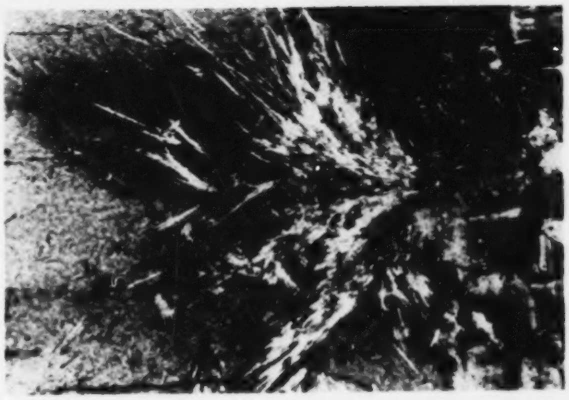


Figure 9.42. Photomicrograph of fibrolite in altered felsic metavolcanic rocks in the Marshall Lake area. Magnification 12.5x, crossed polars (photograph *courtesy of* S.E. Amukun).

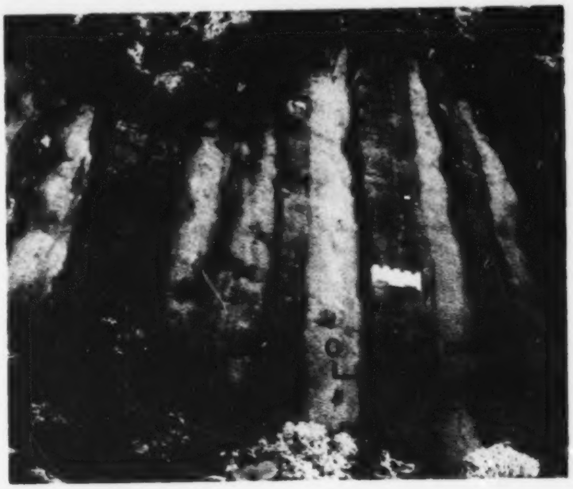


Figure 9.43. Interbedded magnetite iron formation and wacke in the Bending Lake area, south of Dryden (photograph *courtesy of* C. Blackburn).

Rainy Lake

Many copper-nickel and iron occurrences are associated with mafic to ultramafic intrusions in the Rainy Lake area (Poulsen 1984a). The occurrences are mainly within the Grassy Portage and Bad Vermilion Lake intrusions. Poulsen (1984a) has classified mineralization in these intrusions into: 1) copper-nickel mineralization at the base of the Grassy Portage intrusion; 2) copper mineralization near the top of the Grassy Portage intrusion; and 3) iron-titanium mineralization in the upper-levels of both intrusions.

The best example of type 1 is the Northrock deposit where up to 1.0 Mt grading 1.17% Cu (Poulsen 1984a) comprises disseminated chalcopyrite and pyrrhotite with pentlandite and pyrite, in gabbro, metagabbro and leucogabbro. Mineralization forms 3 *en-échelon* lenses near the base of the Grassy Portage intrusion (Harris 1974).

An example of the second type is the Mironsky showing where disseminated pyrrhotite-chalcopyrite mineralization is related to siliceous zones near the top of the Grassy Portage intrusions. Approximately 300 000 tons of material grading 0.8% Cu are present (Harris 1974). The mineralization may represent granophyric differentiates, or mineralized assimilated blocks (Poulson 1984a).

The third type consists of lenticular zones of disseminated to massive magnetite and ilmenite with apatite and local rutile in the central to upper levels of both the Grassy Portage and Bad Vermilion intrusions. Up to 2 Mt of magnetite-ilmenite mineralization grading up to 45% combined Fe-TiO_2 was outlined by diamond drilling north of Seine Bay (Poulsen 1984a).

Chromite

A number of chromite occurrences are scattered throughout the Wabigoon Subprovince and are associated with ultramafic intrusions. Minor chromite occurs as disseminated grains and layers up to 1 cm wide in the northern ultramafic part of the Lac des Iles intrusive complex (Sutcliffe et al. 1989). A small deposit occurs at the northern margin of the Obonga Lake greenstone belt (Whittaker 1986). A shaft was sunk to 110 m by Chromium Mining and Smelting Corporation Limited in the 1930s and 25 kt averaging 12% Cr was outlined by diamond drilling (Whittaker 1986).

Epigenetic Mineral Deposits

Some degree of structural control on emplacement has probably been effective for all classes of mineral deposits in the Wabigoon Subprovince. The leading example is gold, where a number of workers have emphasized the location of gold deposits relative to particular rock types (Hurst 1933; Thomson 1936b; Moore 1940; Riley et al. 1971; Mackasey et al. 1974; Blackburn and Janes 1983; Melling et al. 1988). Compilations of deposit locations (Riley et al. 1971; Mackasey et al. 1974; Blackburn and Janes 1983) indicate their presence in all broad categories of Archean rocks. This fact is a strong argument that the final concentrating process was structurally controlled and was Neoarchean in age.

GOLD MINERALIZATION

Age determinations from the Duport Mine at Shoal Lake constrains gold deposition to between the age of 2716±2 Ma on a mineralized dike in the deformation zone and a dike which crosscuts mineralization dated at 2709±2 Ma (Davis and Smith 1991). These data may not be diagnostic of all deposits but agree well with observations made by Colvine et al. (1988, p.59) that for all Ontario Superior Province deposits "gold introduction occurred during the latest Archean, and at essentially the same time across extremely large areas of the Superior Province."

At Shoal Lake, gold mineralization at the Duport Mine, is contained within the steep west-dipping, northeast-trending Duport deformation zone. The zone lies parallel to, but locally crosscuts, pillowed volcanic rocks and dioritic intrusions. In the vicinity of the gold mineralization, the deformation zone contains a retrograde, greenschist-facies mineral assemblage which overprints an amphibolite-facies metamorphic aureole of the adjacent Snowshoe Bay batholith. Smith (1986) interpreted the reverse sense of movement across the sheared zone to have been due to emplacement of the Snowshoe Bay batholith. Gold mineralization, accompanied by sulphidization, biotitization, silicification and carbonatization, is predominantly of 2 kinds: within a layered sulphide-quartz rock; and within a siliceous breccia in which sulphides are concentrated with quartz along margins of the green, micaceous fragments.

A second example of a late-emplaced deposit is the Cameron Lake gold deposit, situated in the Cameron Lake volcanics in the Kakagi–Rowan lakes greenstone belt, which has been demonstrated (Melling et al. 1986) to be a structurally controlled, quartz-carbonate breccia vein-type deposit within deformed metabasaltic rocks of the Cameron Lake shear zone. The shear zone is a splay from the Pipestone–Cameron deformation zone, which cuts through the stratigraphic units of volcanic rocks and exhibits evidence of dextral, strike-slip motion.

Sulphide replacement has been demonstrated to be a significant concentrating process, for example at the MacLeod–Cockshutt and Hard Rock mines, in the Geraldton area (Macdonald 1983). Pyritic replacement of magnetite iron formation followed development of a late, regional cleavage along the Wabigoon–Quetico subprovincial boundary and accompanied veining and gold deposition in shear zones. Commonly, iron-bearing carbonate is a second alteration product (Macdonald 1983) closely related to the sulphide replacement, also demonstrated at the Duport (Smith 1986) and Cameron Lake (Melling et al. 1986) deposits.

The role of brittle-ductile deformation in felsic plutons subjected to severe strain has been demonstrated by Poulsen (1984a, 1984b) at Mine Centre. Here, strain taken up largely along the Quetico and Seine River faults was also translated into openings along ductile shear zones and open fractures occupied by gold-bearing quartz veins at high angles to a regional sigmoidal shear pattern. Poulsen (1984a, 1984b) indicated that virtually all quartz shows microscopic evidence of postdepositional deformation at moderate temperatures, resulting in a "cherty", locally laminated appearance to the quartz. Gold is concentrated in this quartz variety, clearly indicating the very close relationship between the gold concentrating process and late, regional transpression at the subprovincial boundary, in this case.

POST-ARCHEAN EVENTS

The Wabigoon subprovince was the site of Neoarchean to Mesoproterozoic magmatism consisting of alkalic complexes and diabase dikes. The alkalic complexes are the Sturgeon Narrows Alkalic Complex (2746 Ma) and the Chipman Lake fenites (1029 Ma) (cf., Sage , this volume). The former is located along an Archean shear zone within the Sturgeon Lake greenstone belt, whereas the latter lies along the Trans–Superior Tectonic Zone. These complexes are discussed in the chapter on alkalic complexes (*see* Sage, this volume). Diabase dikes of the Matachewan (2.45 Ga), Kenora–Fort Frances (2.2 to 2.1 Ga), Marathon (2.17 Ga), Wabigoon (approximately 1.9 Ga) and Nipigon (1.1 Ga) swarms (Fahrig and West 1986) cut rocks of the Wabigoon Subprovince. The characteristics, extent, petrography, geochemistry and tectonic setting of these dike sets are reviewed by Osmani , this volume. The subprovince is also cut by the 1.1 billion-year-old Logan diabase sills and coeval ultramafic intrusions (*see* Sutcliffe, this volume).

Limited investigations within the Wabigoon Subprovince suggest that the subprovince has been affected by Proterozoic compression, forming discordant brittle structures along which late hydrothermal minerals were deposited (Kamineni et al. 1990). In common with most of the Superior Province, Wabigoon subprovince rocks are affected by post-Archean brittle fractures cutting all rock types (evident on all large-scale compilations of lineaments; e.g., Lowman et al. 1987). Kamineni et al. (1990) have observed both transcurrent and thrust movement on some of the fractures. The Eye–Dashwa Pluton has been dated at *circa* 2665 Ma by the U-Pb zircon method (Zartman and Kwak 1990); a whole rock Rb-Sr age of 2637±33 Ma (Kamineni et al. 1990) was also obtained on unaltered parts of the pluton. A younger Rb-Sr age (2280±167 Ma) was obtained on altered parts of the pluton. These results suggest the Rb-Sr ages may be dating Proterozoic alteration events in the pluton (Kamineni et al. 1990).

TECTONIC SYNTHESIS

In this section, we document the criteria for recognition of tectonic assemblages within the 3 major divisions of the Wabigoon Subprovince. A definition of tectonic assemblages is offered in Chapter 4 (*see* Thurston, this volume). This is followed by discussion of a number of aspects of the tectonics of the subprovince including: the relationships of the Wabigoon Subprovince with the neighbouring Winnipeg River and Quetico subprovinces; and tectonic patterns in the Wabigoon Subprovince. A model is put forward to explain the geological data.

Tectonic Assemblages

Definition of tectonic assemblages has been made by compiling stratigraphic-facing determinations, geochronologic and stratigraphic data in order to assess whether homoclinal or folded panels of supracrustal units could be recognized and traced through the subprovince as a whole or any of its constituent regions. The assemblages follow from the time-stratigraphic correlation of the WWR discussed earlier. The recognition of tectonic assemblages admits that not all lithostratigraphic sequences are in chronologic order, some have fault-bounded relationships, and that considerable time has elapsed between the first- and last-formed rocks in the region.

WESTERN WABIGOON

In the WWR, volcanic successions have been assigned to 3 distinct successions, a lower mafic, an overlying diverse intermediate to felsic, and an upper mafic sequence (Trowell et al., 1980; Blackburn et al. 1985). Geochronologic data (*see* Table 9.7) in turn allow a finer subdivision of WWR metavolcanic sequences into chronologic groups: type 1a (2775 Ma); type 1b (2745–2730 Ma); and type 1c (2729–2725 Ma). The results of this process in the WWR are portrayed in Figure 9.44 and described more fully in Table 9.7. Critical points arising from this interpretation are discussed in the following paragraphs.

Assemblages are shown and named only in instances in which geochronologic and stratigraphic control allow demonstration that an assemblage is a geochronologic and stratigraphically consistent entity. In any attempt such as this to erect a large-scale synthesis, individual assemblages will incorporate numerous, slightly differing stratigraphic units. The most important test of a process such as this is whether the assemblages erected produce a large-scale order to the WWR. The following paragraphs are a guide to Table 9.13 and Figure 9.44.

The *circa* 3 billion-year-old Lumby assemblage, which includes the Steep Rock Group of Wilks and Nisbet (1988) plus some adjacent granitoid bodies, is the oldest unit within the WWR. The boundaries of the 3 billion-year-old region are not known. The Fourbay assemblage in the northern part of the Sturgeon Lake area is 2.77 billion years in age. The Fourbay assemblage is disconformably(?) overlain by the 2.74 billion-year-old Handy assemblage which is traceable to the Savant Lake area to the north.

The Wabigoon assemblage composes a number of 2735 million-year-old, predominantly mafic lithostratigraphic units (*see* Table 9.13). The mafic units are here postulated to form a single assemblage by virtue of similar rock types, age and relationships to neighbouring units (*see* Table 9.13). Tracing of inclusion trains of this assemblage across the Indian Lake batholith permit extension of the unit around both flanks of the Atikwa–Lawrence batholith and thence westward to the southern end of the Lake of the Woods area. We do not indicate a distinct, named assemblage comprising the bulk of the supracrustal rocks in the Lake of the Woods area because of the lack of geochronologic control. However, a single age determination of 2727 Ma is available in the High Lake area (Davis and Smith 1991).

The younger (2725 to 2711 million-year-old) Central Sturgeon, Boyer and Kakagi assemblages are each at least partly fault bounded. The Post Lake sediments disconformably overlie the 2736 million-year-old (Davis et al. 1985) Central Sturgeon volcanics (Trowell 1983); the disconformable relationship suggests the Post Lake sediments are possibly type 2 sedimentary rocks (cf., Table 9.13). The Savant

Lake group (Trowell 1986a) and Abram group sedimentary rocks (Turner and Walker 1973) bear similar stratigraphic relations to their respective underlying, largely 2735 to 2733 million-year-old metavolcanic rocks (Davis et al. 1988). The Abram group has been dated at less than 2698 Ma (Davis and et al. 1988). The Warclub group is constrained by the 2714 million-year-old (Davis and Edwards 1982) age of the Berry River formation metavolcanic rocks within the assemblage; therefore, the absolute ages available do not clearly reflect a single age of sedimentation. Units such as the Crowduck Lake group and the White Partridge Bay group form the upper part of supracrustal units in the Lake of the Woods area. As such, they are probably less than 2710 million years in age, but this conclusion is speculative until age determinations are available, as the units vary from pre- to syndeformational in age. The Warclub assemblage at present includes the Minnitaki sediments which form a self-contained unit. These stratigraphic units are grouped into a single assemblage based upon rock type, age and sedimentary and structural relationships to older meta-volcanic assemblages.

CENTRAL WABIGOON

The CWR consists of a basement complex of uncertain extent and greenstone belts ranging in age from the approximately 3 billion-year-old Lumby Lake and Steep Rock Lake belts (Davis and Jackson 1988) to the approximately 2750 million-year-old large rhyolitic unit in the Obonga Lake belt. Many of the greenstone belts may be divizable into separate assemblages once absolute ages and geochemical data are available for them. Subdivision of the plutonic and metamorphic rocks requires mapping and age determinations. In general, the Mesoarchean basement complex postulated to exist in the southern part of the CWR may not extend north as far as Obonga Lake. However, the approximately 3 billion-year-old age of a granitic unit near Armstrong (Davis et al. 1989) suggests much complexity remains to be unravelled.

EASTERN WABIGOON

A small number of assemblages were suggested by Blackburn and Johns (1988) for the much smaller, less-documented EWR. Lack of stratigraphic-facing data, poor knowledge of contact relationships and structural complexities has meant that a similar tripartite division used in the WWR is inappropriate. There is insufficient geochronologic and structural data to support subdivision into tectonic assemblages.

Subprovince Margin Relationships

A full understanding of the geologic development of a granite-greenstone subprovince is in part determined by an understanding of how it relates to its neighbouring subprovinces.

WABIGOON–WINNIPEG RIVER RELATIONSHIPS

Overthrusting of Wabigoon Subprovince supracrustal rocks onto Winnipeg River Subprovince sialic crust was preceded by north-dipping subduction along the southern boundary of the Winnipeg River Subprovince (Davis and Smith 1991). This proposal is supported by zircon inheritance patterns in late plutons from the WWR; the plutons are the only rocks within the WWR showing evidence of zircon inheritance from a source significantly older than volcanism. In the case of the Snowshoe Bay batholith, at Shoal Lake, inheritance prevented precise crystallization age determination, but it is less than 2708±7 million years old, and shows evidence of inheritance from a source at least 2800 million years in age (Davis and Smith 1991). These data led Davis and Smith to suggest that sialic material of the Winnipeg River Subprovince underlies greenstones of the Wabigoon Subprovince at Shoal Lake at 2709 Ma.

WABIGOON–QUETICO RELATIONSHIPS

Along the southern margin of the Wabigoon Subprovince, development of Quetico Subprovince sedimentary rocks as prograding alluvial to submarine fans in an accretionary wedge above a north-dipping subduction zone is supported by structural and sedimentologic evidence (Devaney and Williams 1989; Williams 1990). In the Beardmore–Geraldton Belt, located just north of the Quetico Subprovince, repetitive supracrustal couplets consisting of lower basaltic and upper clastic sedimentary units, have relatively proximal sedimentary facies juxtaposed on top of more distal ones (Devaney and Williams 1989; *see* Williams, this volume).

Geochronologic data, on 26 zircon grains from Quetico metasedimentary rocks just south of the Quetico Fault, near Atikokan, indicate a maximum age for deposition of these sedimentary rocks is less than 2698±3 Ma, the youngest zircon age. Some of the detritus came from terrane as old as 3009±4 million years (Davis et al. 1989), in accord with similar data (range 3.3 to 2.7 Ga) from the Quetico Subprovince (Percival and Sullivan 1988). Ages of the order of 3 Ga have been identified in the adjacent Marmion Lake batholith to the north of the metasedimentary rocks from which the dated zircons came, suggesting that the CWR may have been part of the sedimentary source area. Only 3 grains fell in the range 2750 to 2710 Ma, the period of most intense calc-alkaline volcanism in the Wabigoon Subprovince. The majority of grains only slightly predated the age of sediment deposition; perhaps much of the detritus came from erosion of active felsic volcanic centres and/or unroofing of late plutons, rather than erosion of older volcanic terrane (Davis et al., in press).

Tectonic Patterns within Wabigoon Subprovince

Large-scale tectonic patterns within Wabigoon Subprovince are related to compression at a high angle to the long axis of the subprovince. There is evidence also for a superimposed or concomitant strike-slip component of movement, along discrete zones of deformation, such as the Quetico, Wabigoon, Pipestone–Cameron, Crowduck Lake–Witch Bay, and Paint Lake faults. The predominance of Z- over S-style minor fold asymmetries adjacent to these faults, coupled with the sense of deflection of lithologic units into the fault zones, supports a dextral component of movement that is characteristic of late movement throughout the western Superior Province as a whole. The integration of contemporaneous compression across the subprovince, along with dextral sense strike-slip motion at and along its boundaries, is an example of transpression that is characteristic of modern orogens (Stott 1986; Sanborn-Barrie 1991b).

TECTONIC MODEL

This review of a number of models generated over the past 15 years will show the evolution of geologic thought with regard to the development of Wabigoon Subprovince.

Although lithologic and geochronologic data for the Wabigoon Subprovince are among the most precise and most voluminous for any Archean granite-greenstone belt, there are still sufficient gaps to make any one model equivocal. The WWR was one of the first granite-greenstone terranes to have a mobilist tectonic model developed from the geologic data (Langford and Morin 1976; Blackburn 1980).

The Concept of Arc-Accretion

Precise U-Pb zircon geochronology indicates a southward decrease in subprovince accretion age from Sachigo Subprovince in the north to the Wawa Subprovince in the south (Krogh and Davis 1971; Stott et al. 1987; *see* Corfu and Davis, this volume). The early geochronologic work (Krogh and Davis 1971) and scant geochemical and geophysical data led Langford and Morin (1976) to suggest, by analogy with the Phanerozoic Cordilleran belts of Western Canada, that the subprovinces had evolved by successive accretion of island arcs against a Berens River and Gods Lake Craton. In this model, the English River Subprovince, including the present Winnipeg River Subprovince, initially was viewed as being a fore-arc basin of the Uchi Subprovince. Subsequently, it was considered to be a back-arc basin to the Wabigoon Subprovince, with the Quetico Subprovince being a fore-arc basin to the Wabigoon Subprovince.

Rifting of an Older Continental Mass

Evidence of rifting of an old continental terrane as a precursor to development of Wabigoon Subprovince greenstone sequences is sparse. However, at Savant Lake (Cortis et al. 1988; Sanborn-Barrie 1989) and at Obonga Lake (Cortis et al. 1988), unconformities are overlain by sedimentary rocks that might represent a stable, continental margin sequences. Near the southern margin of the subprovince, the unconformity of the Steep Rock Group over the Marmion Lake batholith is evidence of such a margin: the basal stromatolitic carbonate (Mosher Carbonate) with its irregular, subaerial, karstic top suggests initial shallow-water deposition; the overlying goethite- and hematite-rich breccias and hematitic pisolites (Joliffe Ore Zone Formation) suggest a paleosol; and an extensive dike swarm in the basement rocks which predates the unconformity indicates the development of extensional magmatism prior to rifting. Although the age of the unconformity is unknown, geochronologic data on titanite elsewhere in the basement complex indicate that a metamorphic event at least 2809 my in age occurred within the central part of the Marmion Lake batholith (Davis and Jackson 1988) and may be related to early continental breakup (Davis et al. 1988).

Production of Greenstone Belts

Regional geologic patterns in the western Wabigoon Subprovince led Blackburn (1980) to propose a model specific to the central part of the subprovince, including the Eagle–Wabigoon–Manitou lakes and Abram–Minnitaki belts; one in which these belts were produced by opening and subsequent closure of a micro-ocean. Rifting of an older gneissic terrain, represented in the north by the Winnipeg River Subprovince, and an unknown terrain to the south, perhaps the central Wabigoon region, was proposed to lead to the development of oceanic crust represented, for example, by the Wapageisi group and possibly the Botham Bay group. Progressive closure of the micro-ocean was suggested to have induced island arcs on either side of the ocean, above subduction zones dipping northward in the north and southward in the south toward the ocean margins. The diverse volcanic successions overlying the lower mafic volcanic rocks (e.g., Lower Wabigoon volcanics, Pincher Lake group, Upper Manitou Lake group, Kawashegamuk group, and Neepawa group) would be representative of these island arcs, while sedimentary groups like the Manitou Lakes group, the Minnitaki group, and the Abram and Patara groups would represent either fore- or back-arc basin accumulations. Upon closure of the micro-ocean, thrust faulting involving all these volcanic and sedimentary sequences, followed by rotation to a vertical position, was seen to lead to the complex stacking observed today.

Subsequent to the hypothesis of Blackburn (1980), geochronologic work has confirmed the presence of age anomalies that may be interpreted as a result of thrusting, both in the Abram and Minnitaki lakes greenstone belts where the older Neepawa group structurally overlies the younger Abram group, and in the southern part of the Eagle–Wabigoon–Manitou Lakes greenstone belt where the older Boyer Lake group structurally overlies the younger Manitou and Stormy groups. Thrusting along the boundary between the Wabigoon and English River–Winnipeg River subprovinces, along with other evidence, is indicative of arc-continent collision in the period 2709 to less than 2698 Ma (Davis et al. 1988). The Abram and Minnitaki group sedimentary rocks may have received detrital input from a relatively old, continental land mass as well as from plutons unroofed within the overthrusting slab. This proposal is supported by geochronologic data from 9 single zircon grains (D.W. Davis, Royal Ontario Museum, personal communication, 1990) from the Warclub group near Dryden (correlative with the Minnitaki group). Eight of the 9 grains are older than any volcanic rocks in the Wabigoon Subprovince yet dated in this marginal area, the oldest being 3001 million

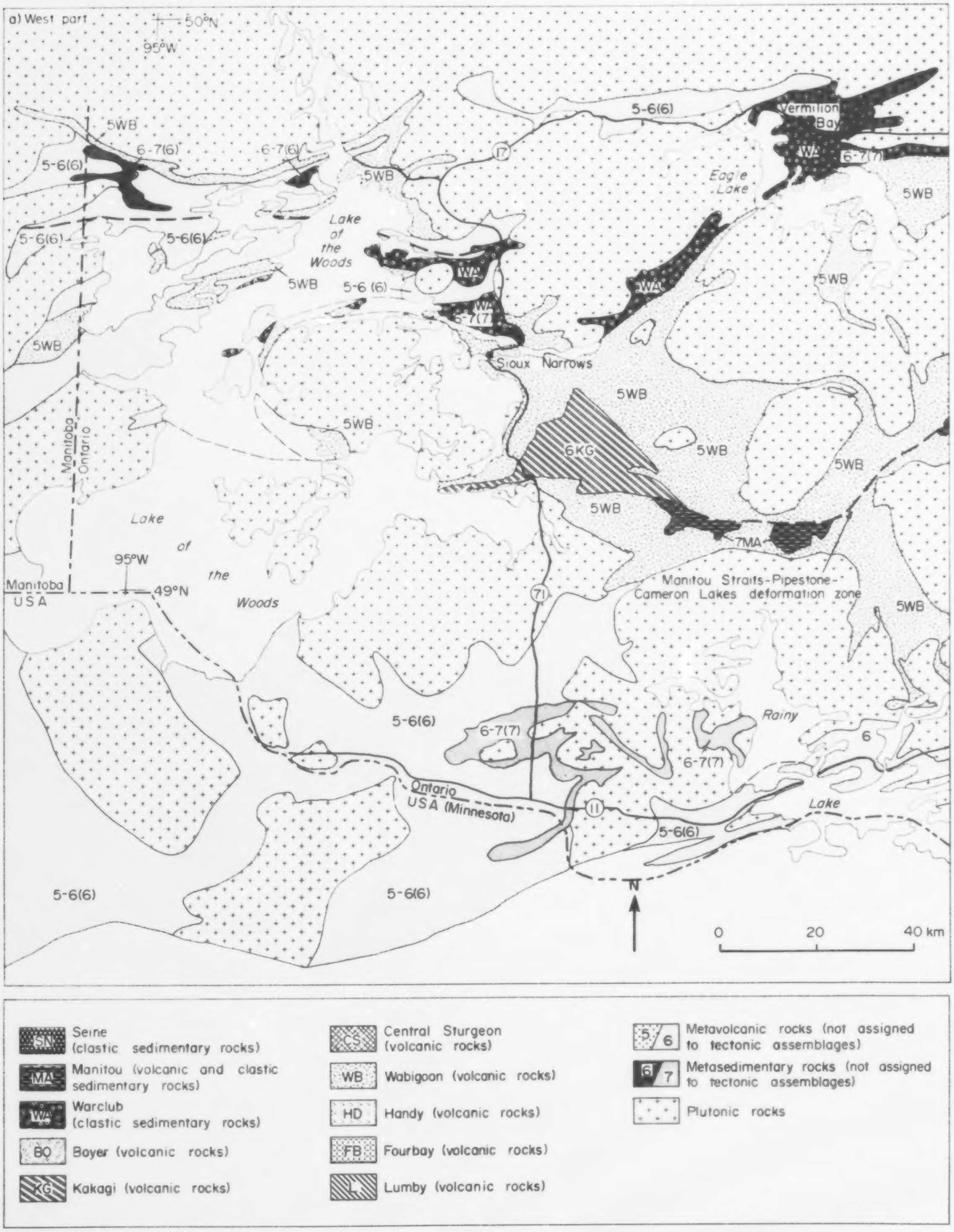

a) West part
50°N
95°W
5WB
6-7(6)
5-6(6)
6-7(6)
5WB
17
5-6(6)
Vermilion Bay
WA
6-7(7)
Eagle Lake
5WB
Lake of the Woods
5-6(6)
5-6(6)
WA
5WB
5-6 (6)
WA
6-7(7)
WA
5WB
5WB
Sioux Narrows
5WB
5WB
Manitoba
Ontario
6KG
5WB
5WB
5WB
Lake of the Woods
95°W
7MA
Manitou Straits-Pipestone-Cameron Lakes deformation zone
Manitoba
USA
49°N
71
5WB
5-6(6)
6-7(7)
Rainy
6
6-7(7)
Ontario
USA (Minnesota)
11
Lake
5-6(6)
5-6(6)
5-6(6)
N
0
20
40 km
SN Seine (clastic sedimentary rocks)
MA Manitou (volcanic and clastic sedimentary rocks)
WA Warclub (clastic sedimentary rocks)
BO Boyer (volcanic rocks)
KG Kakagi (volcanic rocks)
CS Central Sturgeon (volcanic rocks)
WB Wabigoon (volcanic rocks)
HD Handy (volcanic rocks)
FB Fourbay (volcanic rocks)
L Lumby (volcanic rocks)
5/6 Metavolcanic rocks (not assigned to tectonic assemblages)
6/7 Metasedimentary rocks (not assigned to tectonic assemblages)
Plutonic rocks

Figure 9.44. Tectonic assemblage interpretation, Wabigoon Subprovince. Letter codes correspond to those found on the Tectonic Assemblages of Ontario (*see* Maps 2575 to 2577, map case). Also see Maps 2575 to 2577 for an explanation of numerical codes.

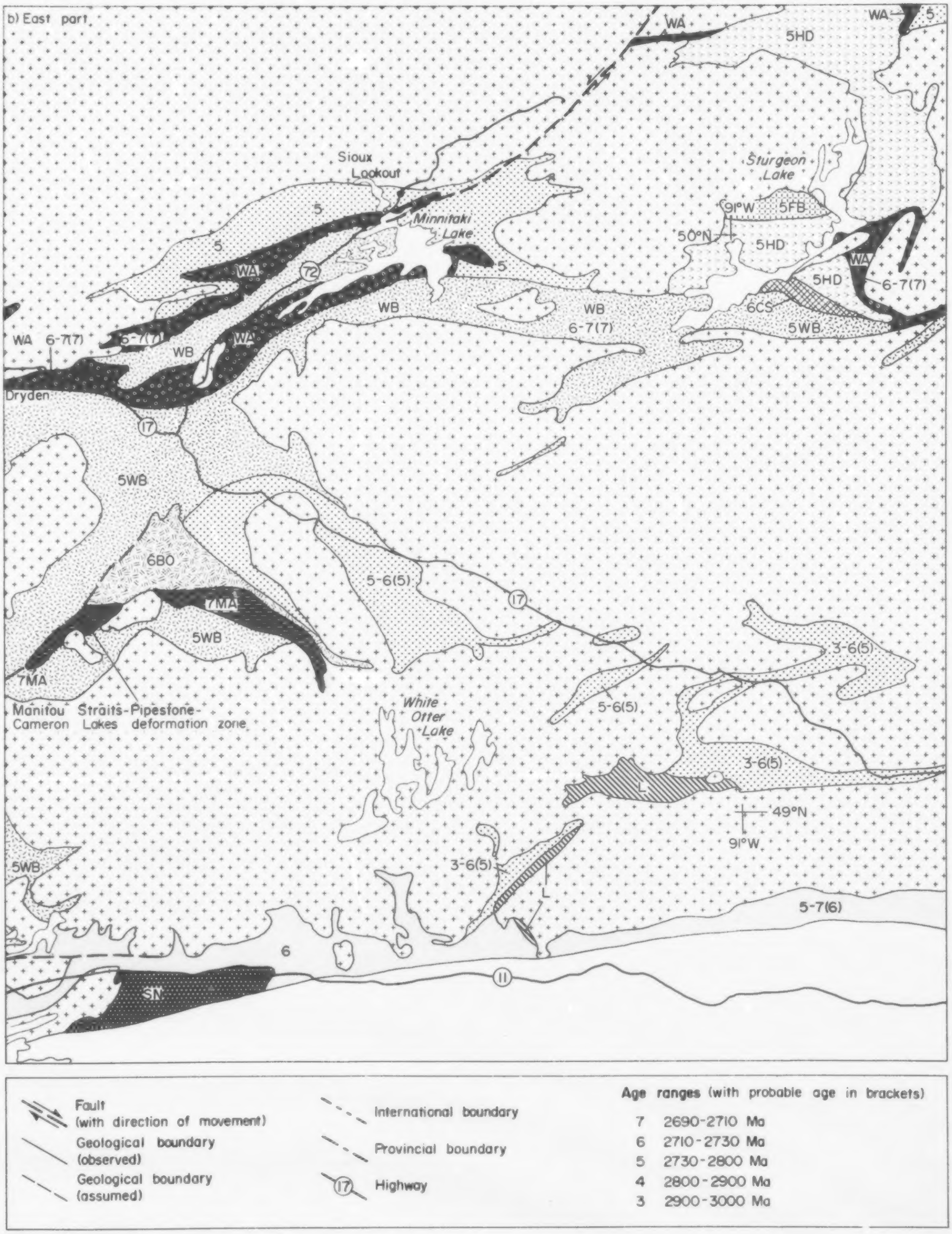

b) East part
WA
5
5HD
Sioux
Lookout
Sturgeon
Lake
91°W
5FB
50°N
5HD
Minnitaki
Lake
5
5
5
WA
72
5
WA
5HD
6-7(7)
6CS
5WB
WB
WB
6-7(7)
WA 6-7(7)
6-7(7)
WB
WA
Dryden
17
5WB
6BO
5-6(5)
17
7MA
5WB
7MA
3-6(5)
Manitou Straits-Pipestone-
Cameron Lakes deformation zone
White
Otter
Lake
5-6(5)
3-6(5)
L
49°N
91°W
5WB
3-6(5)
L
5-7(6)
6
11
SN
Fault
(with direction of movement)
Geological boundary
(observed)
Geological boundary
(assumed)
International boundary
Provincial boundary
17 Highway
Age ranges (with probable age in brackets)
7 2690-2710 Ma
6 2710-2730 Ma
5 2730-2800 Ma
4 2800-2900 Ma
3 2900-3000 Ma

Figure 9.44. Tectonic assemblage interpretation, Wabigoon Subprovince. Letter codes correspond to those found on the Tectonic Assemblages of Ontario (*see* Maps 2575 to 2577, map case). Also see Maps 2575 to 2577 for an explanation of numerical codes.

Table 9.13. Examples of tectonic assemblages and their constituent stratigraphic units.

Type of Assemblage and Examples	Examples of Stratigraphic Units	Contact Relationships with Adjacent Assemblages	Comments
Type I Extensional environment on continental crust			
i) Lumby assemblage (3 Ga)	Steep Rock Group—platformal sedimentary and volcanic sequence	unconformably overlying 3.0 billion-year-old Marmion Lake batholith	may be part of an older microcontinent, the remnants of which may be found in small greenstone belts in central Wabigoon Subprovince and north of Lake Nipigon
Type II oceanic island, arc volcanism			
i) Fourbay assemblage (2775 Ma)	North Sturgeon Lake volcanics–tholeiitic flows and thin units of calc-alkalic pyroclastic rocks	speculative disconformable boundary with overlying 2745 million-year-old Handy volcanics	basaltic assemblage tentatively based on significant 30 million year difference in age of volcanism from overlying Handy assemblage
ii) Handy volcanics (2745 Ma)	Handy volcanics: tholeiitic flows and calc-alkalic pyroclastic rocks	unconformably overlain by sedimentary rocks (Savant Lake group)	
iii) Upper and Lower Wabigoon volcanics (2736 to 2732 Ma)	Wapageisi group and Katimiagamak group: dominantly magnesian tholeiitic flows; Upper Manitou Lake volcanics, Upper and Lower Wabigoon volcanics: tholeiitic to calc–alkalic flows and pyroclastic rocks	fault contact with Boyer, Central Sturgeon, Warclub and Manitou assemblages; speculative faults or unconformities at contacts with other assemblages	Neepawa volcanic rocks probably an allochthonous slice derived and transported from the main mass of Wabigoon assemblage; multiple thrust stacking of Sturgeon Lake volcanic strata
iv) Kakagi Lake group (2725 to 2711 Ma)	Kakagi Lake group: calc-alkalic pyroclastic rocks	possible fault contact with westward extension of Manitou assemblage.	
v) Boyer Lake group (2722 Ma)	Boyer Lake group: dominantly iron-rich tholeiitic flows.	fault contact with Manitou and Wabigoon assemblages; uncertain contact to northeast with Kawashegamuk group	assemblage shape, age and fault contacts with adjacent assemblages are consistent with allochthonous interpretation
vi) Central Sturgeon Lake volcanics (2718 Ma)	iron–rich tholeiitic flows and overlying calc-alkalic pyroclastic rocks	fault contact with Wabigoon assemblage disconformable contact with Post Lake sediments	probably tectonically interleaved between Post lake sediments of Warclub asssemblage and 2736 to 2733 million-year-old Sturgeon Lake volcanic rocks; sedimentary assemblages may encompass several tectonic environments and have a protracted history of sedimentation prior to and during the *circa* 2.7 billion-year-old orogeny
Type III Synorogenic sedimentation			
i) Warclub clastic sedimentary rocks	Warclub, Minnitaki, Abram, Savant Lake groups and Post lake sediments: alluvial fan-fluvial and resedimented facies; pyroclastic rocks	faults and unconformities against adjacent volcanic assemblages	probably reflects synorogenic erosion and tectonic interleaving with volcanic assemblages
ii) Manitou volcanic and clastic sedimentary rocks	Manitou Lakes group	disconformable to marked angular unconformity with underlying Wabigoon assemblage (Wapageisi group and Katimiagamak group); fault contact with overlying Boyer assemblage	tectonically interleaved with adjacent volcanic assemblages; may have been tectonically transported northward
iii) Seine clastic sedimentary rocks	Seine metasediments: alluvial fan-fluvial facies association	unconformably overlies volcanic units along southern margin of Wabigoon Subprovince; fault contact with Quetico sedimentary assemblage	autochthonous assemblage formed along the southern flank of the Wabigoon Subprovince during the terminating stages of orogeny

years old. The ninth and youngest grain is 2729 million years old.

CONCLUSIONS

It is now generally accepted that the major calc-alkalic sequences within Superior Province greenstone belts formed largely in island arc-related environments (Thurston and Chivers 1990; Card 1990). Arc magmatism is the result of oceanic subduction and may develop on either continental or oceanic crust and commonly in extensional environments associated with back-arc basins. During an advanced stage of ocean closure, subduction of young, relatively buoyant oceanic lithosphere may cause the tectonic regime to become compressional, leading to development of a fold-thrust belt. Eventual complete ocean closure and collision with older continental crust results in intense deformation, uplift and sedimentation.

The pattern of ages in the WWR broadly supports this model. An early extended period of largely oceanic volcanism in the time period 2.78 to 2.72 Ga is followed abruptly at about 2.71 to 2.684 Ga by widespread deformation and deposition of most of the Timiskaming-type and accretionary prism sedimentary units. These metasedimentary rocks contain abundant detritus eroded from the older 3.0 to 2.7 billion-year-old rocks in the adjacent metaplutonic Winnipeg River Subprovince and the CWR. Late plutonic and volcanic rocks in this age range also contain greater than 2.8 billion-year-old zircon inheritance. However, the greater than 2.71 billion-year-old volcanic sequences within the greenstone belt have thus far shown no evidence for interaction with older terranes. Out-of-sequence age relations within internally concordant supracrustal sequences have been found at several locations within the WWR and are probably caused by a late 2.70 billion-year-old period of thrusting which predates isoclinal folding (Davis et al. 1988). Such relations have also been documented at other places in the Superior Province (e.g., Stott, this volume; *see* Jackson and Fyon, this volume; Corfu and Ayres 1991).

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Chapter 10

Quetico Subprovince

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Abstract

The Quetico Subprovince consists predominantly of metamorphosed turbiditic wacke, largely derived from, and deposited during and after, the volcanic climax in the neighbouring Wawa, Wabigoon and Abitibi subprovinces, during the period 2.70 to 2.69 Ga. Overthrust imbrication at both the Wabigoon–Quetico and the Quetico–Wawa subprovince contacts occurred along north-dipping shears, which subsequently steepened to near vertical during compressional tectonics. A regional schistosity, on which fold axes and mineral elongation lineations are variably developed and oriented, is interpreted to result from regional shortening and dextral-sense shearing, especially at the margins of the Quetico Subprovince. Amphibolite-facies metamorphism, migmatite formation and granite intrusion occurred during the period 2.67 to 2.65 Ga.

The Quetico Subprovince may represent a fore-arc accretionary prism.

INTRODUCTION

The Quetico Subprovince, shown in Figures 10.1 and 10.2 , extends from Minnesota in the southwest (Southwick 1972; Bauer 1985, 1986), across Ontario (Lawson 1913; Percival 1989) eastward for nearly 1000 km into Quebec (Lacroix 1987). East of the Kapuskasing Structural Zone (Figure 10.2c), the Opatica Subprovince, as defined by Card and Ciesielski (1986), is an eastward continuation of the Quetico Subprovince. There, in eastern Ontario, metasedimentary and plutonic igneous rocks constitute an extremely poorly exposed belt over 100 km wide having irregular and ill-defined boundaries with the neighbouring subprovinces. In Quebec, usage of the term Opatica Subprovince includes a southern, mainly plutonic igneous suite and a northern metasedimentary belt which is probably equivalent to the Quetico Subprovince (Lacroix 1987). Subdivision and clarification of the relationships between the Opatica Subprovince in both Ontario and Quebec, and the Quetico Subprovince in Ontario, is strikingly similar to the relationships between the English River and Winnipeg River subprovinces (*see* Breaks, this volume).

The Quetico Subprovince is a linear strip of dominantly metasedimentary rocks, and migmatitic and anatectic derivatives that has a relatively consistent width of about 70 km. In general, the boundaries of the Quetico Subprovince are mapped as steeply dipping surfaces across which there is commonly a distinct contrast in rock type (e.g., Mackasey et al. 1974; Carter 1984, 1985, 1986, 1987). For example, in Ontario, the northern boundaries of the Quetico and Opatica subprovinces against the Wabigoon Subprovince, and the southern boundaries of the Quetico and Opatica subprovinces against the Abitibi and Wawa subprovinces have been defined to exclude mafic metavolcanic and derived migmatite units that are typical of all these subprovinces (Sutcliffe 1981, 1986; Blackburn and Mackasey 1977). The subprovince boundaries are presently considered to be predominantly tectonic but in some places may originally have been depositional. For example, in the Shebandowan area, the Wawa–Quetico subprovince boundary has been considered alternatively to be a simple unconformity between northward-younging sedimentary rocks of the Quetico Subprovince and overlying volcanic rocks of the Wawa Subprovince (Harris 1970), or a thrust fault located along a lithologic contact (Perdue 1938; Giblin 1964). Regional-scale evidence (Williams 1990) of the latter alternative is presented and discussed in this chapter. The subprovince boundaries are defined lithologically; however, some are transitional in nature and variable along strike, and are therefore not able to be depicted precisely on a geological map. Elsewhere, the subprovince is bounded by major zones of transcurrent motion, such as the Quetico and Gravel River faults (Card 1982, 1983; Card et al. 1981), or by intrusive granitoid rocks such as the Hopper Lake granitoid complex (*see* Figure 10.2c, Location 2; Marmont 1987). After the work of Percival (1989), continuity of the Quetico Subprovince across the Kapuskasing Structural Zone has been assumed. However, further work is required to characterize the dominantly plutonic nature of the poorly exposed rocks, east of the Kapuskasing Structural Zone, north and northeast of Cochrane, which appear to have much in common with the Opatica Subprovince that is better known from Quebec.

The Quetico Subprovince has been the subject of a review by Percival (1989). Models for its tectonic development and relationships with adjacent subprovinces have been published recently (Percival and Williams 1989; Devaney and Williams 1989; Williams 1989, 1990). These articles document previous work, describe a model for tectonic development of the Quetico Subprovince as a fore-arc accretionary prism and discuss the implications of this model.

This chapter reviews the compositional, structural and metamorphic character of the Quetico Subprovince using data obtained mainly from the region west of Long Lake (*see* Figure 10.2a and 10.2b), documented by Percival (1989) and Williams (1989, 1990); these data are then placed within a geochronological framework. There are

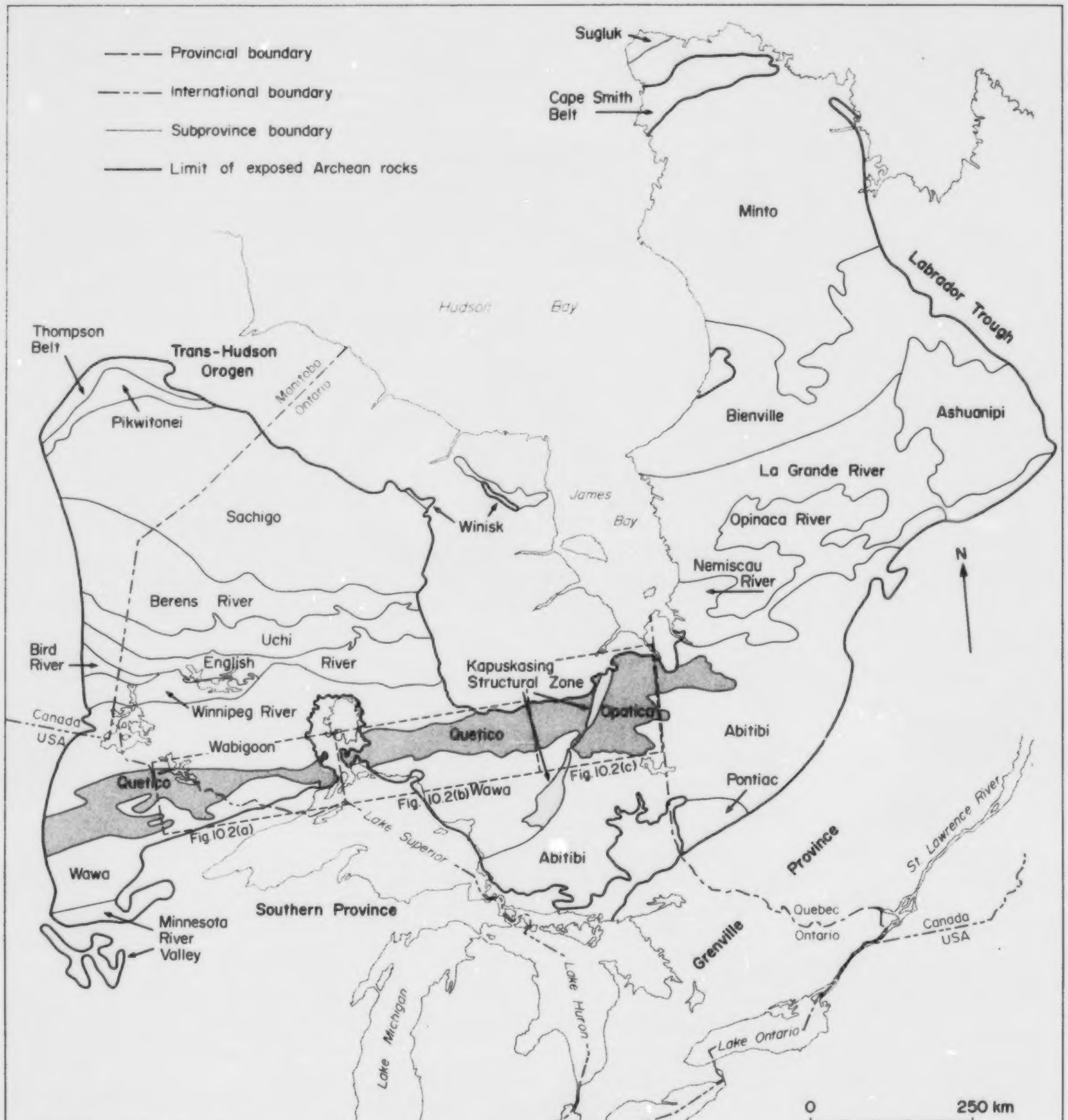

Provincial boundary
International boundary
Subprovince boundary
Limit of exposed Archean rocks
Sugluk
Cape Smith Belt
Minto
Labrador Trough
Hudson Bay
Thompson Belt
Trans-Hudson Orogen
Manitoba
Ontario
Pikwitonei
Bienville
Ashuanipi
La Grande River
James Bay
Sachigo
Winisk
Opinaca River
Nemiscau River
N
Berens River
Uchi
Bird River
English River
Kapuskasing Structural Zone
Opatica
Winnipeg River
Canada
USA
Wabigoon
Quetico
Abitibi
Quetico
Fig. 10.2(c)
Pontiac
Fig. 10.2(b)
Wawa
Fig 10.2(a)
Lake Superior
Abitibi
Wawa
Province
St. Lawrence River
Southern Province
Minnesota River Valley
Quebec
Ontario
Canada
USA
Grenville
Lake Huron
Lake Michigan
Lake Ontario
0
250 km
Lake Erie
Quetico and Opatica Subprovinces

Figure 10.1. Map showing the Quetico Subprovince and its neighbours, the Wawa, Abitibi and Wabigoon subprovinces. Shown, by means of boxes, are the areas covered by the 3 parts of Figure 10.2 (*modified from* Card and Ciesielski 1986).

insufficient data to subdivide the siliciclastic wackes and migmatites into assemblages as has been done within the greenstone-dominated subprovinces. Field evidence of the relationships of the Quetico Subprovince with the adjacent Wabigoon, Abitibi and Wawa subprovinces is documented and the tectonic implications are briefly reviewed.

LITHOLOGIC UNITS

Most of the metasedimentary rocks in the Quetico Subprovince, originally consisted of wacke and siltstone; rare iron formation, conglomerate and metasedimentary rocks of ultramafic composition also occur but are volumetrically insignificant. Igneous rocks include biotite-hornblende-magnetite-bearing granitoid (I-type) bodies of mixed mafic and felsic composition with minor associated ultramafic units; and metaluminous to peraluminous, one- and two-mica leucogranitoids, some of which have a metasedimentary origin (S-type). Classification of granitoid rocks into I- and S-types, representing magmas of mantle or lower crustal and sediment derivation respectively, follows that developed by White and Chappell (1983). Ultramafic bodies, though volumetrically insignificant, are described in detail because these bodies feature in the development of tectonic models for sedimentary subprovinces (Williams 1987b, 1988, 1989, 1990).

Mafic, ultramafic and some felsic units can be distinguished aeromagnetically from the surrounding metasedimentary rocks. Contoured aeromagnetic data allow distinction of areas of more magnetic rock that display regional elongate patterns along the east-northeastward trend of the subprovince (*see* maps 2585 and 2586, map case). Some of these elongate patterns are underlain by small patches of gabbro, or discrete magnetite-bearing felsic intrusions, while others, especially along the southern boundary of the Quetico Subprovince, may be due to high-grade metamorphic assemblages that include magnetic minerals (Williams and Breaks 1989). Few of these anomalies are due to regionally or locally mappable, homogeneous lithologic contrasts; some are due to sheet-like felsic layers up to 1 km in thickness, while others consist of magnetic intrusions interlayered on a smaller scale with the enclosing metasedimentary rocks and derived migmatites. Bouguer gravity anomalies tend to be low and positive, with no steep gradients. A slight positive increase in the Bouguer values occurs at, and within, the southern side of the Quetico Subprovince over much of its length.

Visual distinction of metasedimentary rocks from leucogranites, using Landsat data, has been unsuccessful; leucogranitic bodies are also aeromagnetically indistinguishable from migmatized metasedimentary rocks; contacts on the bedrock geology maps of Ontario (*see* Maps 2542 and 2543, map case) between metasedimentary migmatites and granitoids are interpreted from woefully sporadic data. The following rock descriptions are mainly based on reconnaissance and detailed work in the central and western parts of the Quetico Subprovince (e.g., Sawyer 1983; Percival 1989; Williams 1989). Very little work has been carried out in the central and eastern parts of the Quetico Subprovince, east of Long Lake, by reason of its poor exposure and accessibility (e.g., Bennett et al. 1967; Berger 1986; Coates 1968, 1970; Percival 1985); however, given the limited data available, the character of the Quetico Subprovince is apparently uniform as far east as the Kapuskasing Structural Zone. To the east of the Kapuskasing Structural Zone, the Opatica Subprovince in Ontario, as defined by Card and Ciesielski (1986), may be equivalent to the Quetico Subprovince, but data are too sparse for unequivocal comparison.

Metasedimentary Rocks

Most geological reports concerned with Quetico Subprovince rocks contain only brief descriptions of monotonous wackes containing rare amounts of iron formation and conglomerate (e.g., Fumerton 1986; Harris 1974; Shklanka 1972). Detailed published studies on the metasedimentary rocks of the Quetico Subprovince within Ontario and Minnesota by Pettijohn (1943), Blackburn et al. (1985), Ojakangas (1985), and Devaney and Williams (1989), are the major sources of detailed facies description and analysis. All sedimentary rocks are metamorphosed and the prefix "meta" is implied.

Estimate of original sediment thickness within the sedimentary basin that became the Quetico Subprovince is complicated by insufficient knowledge of the structural geometry across much of the subprovince (*see* a commentary on original sediment thickness *in* Tectonic Interpretation and Summary) .

Four major lithotypes are present: wacke, iron formation, conglomerate, and ultramafic wacke and siltstone. Wacke interbedded with mudstone, prior to metamorphism, consisted of metre- to decimetre-thick alternations of graded and ungraded lithic to feldspathic arenite and siltstone. Similarly, iron formation consisted of decimetre- to centimetre-thick laminated chert-magnetite and chert-magnetite-mudstone rocks. Conglomerate consisted of layers up to 5 m thick, containing volcanic clasts in a sandy to silty matrix. Ultramafic sedimentary rocks were originally of fine grained, serpentinous composition.

The wackes are bedded sandy to silty rocks that represent the predominant rock type, in various states of metamorphism; most were apparently deposited in deep water as unchannelled turbidite fans, based on the bed thickness and grain-size data (P.W. Fralick, Lakehead University, personal communication). In rare instances, wackes of ultramafic composition exhibit climbing cross-stratification and local bedding discordances that indicate frequent and strong current activity, this implies sediment deposition in shallow water.

WACKES AND PARAGNEISSES

The Quetico Subprovince is overwhelmingly composed of grey- to buff-coloured, metre- to decimetre-bedded (Figure 10.3a), variably tectonized sequences of feldspathic, graded and nongraded wackes. Quartz arenites are rare; thin section analysis indicates that feldspar, rock fragments and phyllosilicates commonly account for half the volume of the rock. Loading and dewatering structures, scours, grain-size

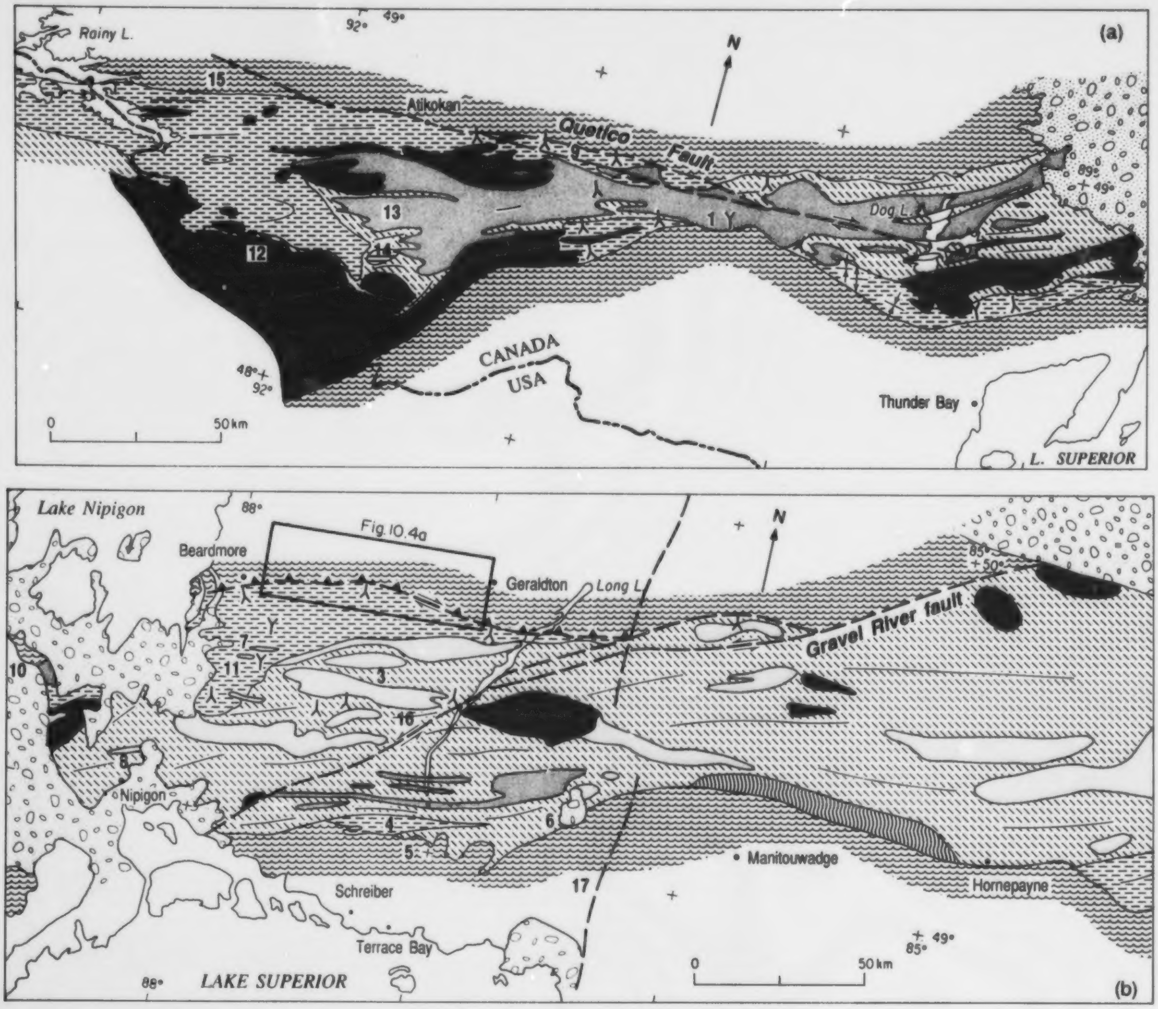

Rainy L.
92°
49°
(a)
N
15
Atikokan
Quetico Fault
13
12
14
1
Dog L.
89°
49°
CANADA
USA
48°
92°
0
50 km
Thunder Bay
L. SUPERIOR
Lake Nipigon
88°
Beardmore
Fig. 10.4a
Geraldton
Long L.
N
85°
50°
Gravel River fault
7
11
3
10
16
8
Nipigon
4
5
6
Manitouwadge
Hornepayne
Schreiber
17
Terrace Bay
88°
LAKE SUPERIOR
0
50 km
85°
49°
(b)

Figure 10.2. Simplified geological and tectonic map of the Quetico Subprovince indicating the features and places described and referenced in the text. For positions of the 3 sections, refer to Figure 10.1.

grading, rare ripple marks, cross-stratification and local discordances indicate that the sedimentary rocks regionally young northwards; although younging reversals are common on a metre- to kilometre-scale. Bedded units are rarely more than 1 m thick, consisting of a micaceous arenite base, rarely conglomeratic, which commonly grades into a featureless, rarely laminated, locally cross-stratified zone capped with a fine-grained top (Figures 10.3a and 10.3b). These compositional types are stacked in upwardly fining Bouma sequences (P.W. Fralick, Lakehead University, personal communication). The bases of many units exhibit loading and convoluted sections, which with abundant flame structures, attest to the rapid sedimentation and dewatering of these rocks. Dewatering of some units was sufficiently vigorous locally to produce decimetre-scale sandstone balls and chaotic breccia-like structures in a homogeneous pelitic matrix, and in many cases, obliterates evidence of internal bedding structures (*see* Figure 10.3c). Rarely, beds several metres thick contain internal scours, little evidence of grading and display a cross-laminated top.

Rare examples of complete beds containing angular to rounded fragments, up to several metres across, are interpreted as disruption of bedding by gravity sliding. Sediment fragments forming a restricted layer near the base of a bed are considered to be an intraformational conglomerate. Both processes, slumping and cannibalization, occurred prior to folding and schistosity formation.

Postcleavage concordant hydrofracture breccias are reported by Sawyer (1984), and are similar to ones documented by Williams (1987a) in which the time relationship between breccia formation and cleavage development is not clear.

The wacke lithofacies contains graded and ungraded wacke beds (Williams 1987a, 1989). There is no documentation of regional or local evidence showing consistent

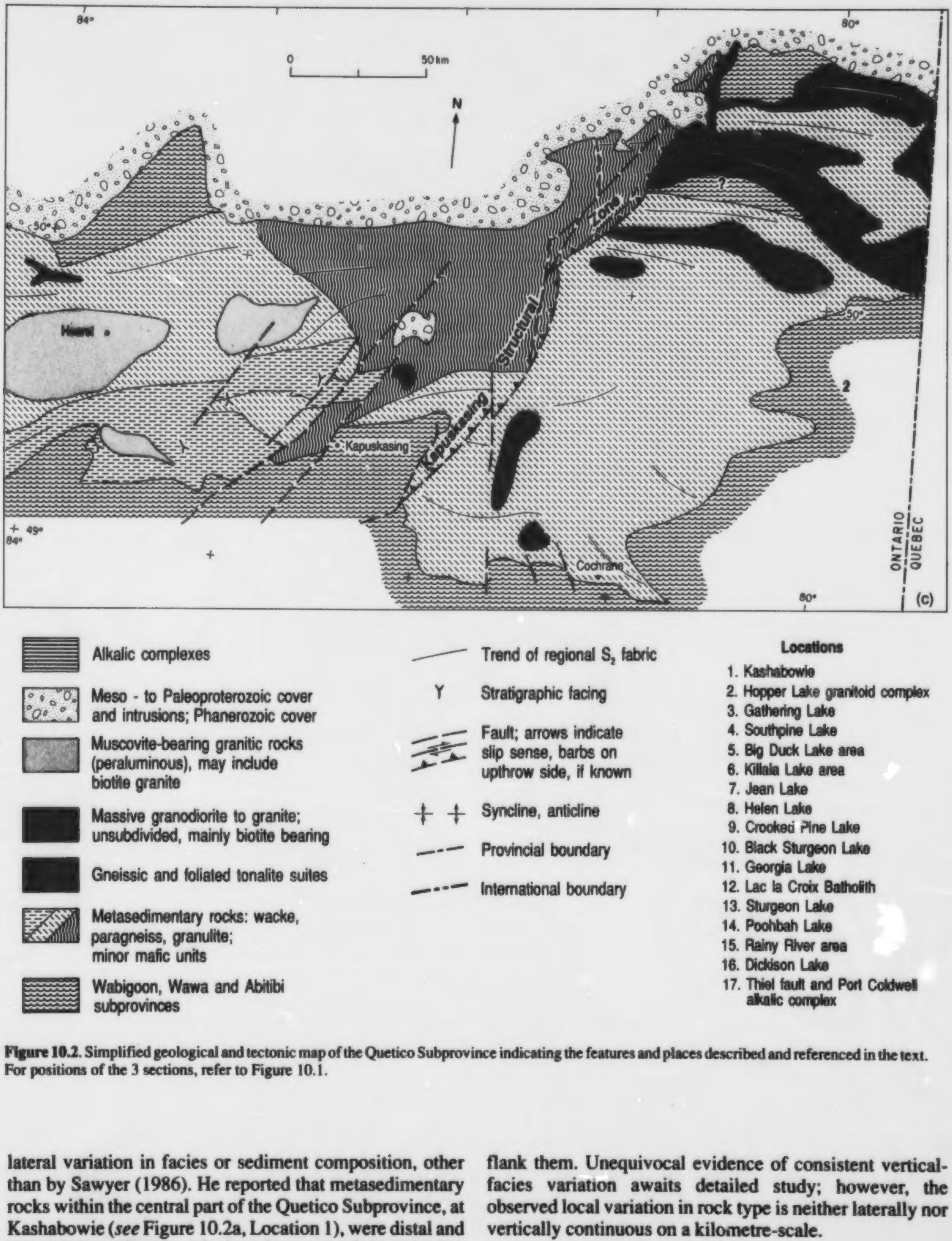

84°
0
50 km
N
80°
50°
Hearst
?
Structural
Zone
Kapuskasing
Kapuskasing
Cochrane
2
49°
84°
80°
ONTARIO
QUEBEC
(c)
Alkalic complexes
Meso - to Paleoproterozoic cover and intrusions; Phanerozoic cover
Muscovite-bearing granitic rocks (peraluminous), may include biotite granite
Massive granodiorite to granite; unsubdivided, mainly biotite bearing
Gneissic and foliated tonalite suites
Metasedimentary rocks: wacke, paragneiss, granulite; minor mafic units
Wabigoon, Wawa and Abitibi subprovinces
Trend of regional S_2 fabric
Y Stratigraphic facing
Fault; arrows indicate slip sense, barbs on upthrow side, if known
Syncline, anticline
Provincial boundary
International boundary
Locations
1. Kashabowie
2. Hopper Lake granitoid complex
3. Gathering Lake
4. Southpine Lake
5. Big Duck Lake area
6. Killala Lake area
7. Jean Lake
8. Helen Lake
9. Crooked Pine Lake
10. Black Sturgeon Lake
11. Georgia Lake
12. Lac la Croix Batholith
13. Sturgeon Lake
14. Poohbah Lake
15. Rainy River area
16. Dickison Lake
17. Thiel fault and Port Coldwell alkalic complex

Figure 10.2. Simplified geological and tectonic map of the Quetico Subprovince indicating the features and places described and referenced in the text. For positions of the 3 sections, refer to Figure 10.1.

lateral variation in facies or sediment composition, other than by Sawyer (1986). He reported that metasedimentary rocks within the central part of the Quetico Subprovince, at Kashabowie (*see* Figure 10.2a, Location 1), were distal and less rich in mafic minerals than the proximal deposits that flank them. Unequivocal evidence of consistent vertical-facies variation awaits detailed study; however, the observed local variation in rock type is neither laterally nor vertically continuous on a kilometre-scale.

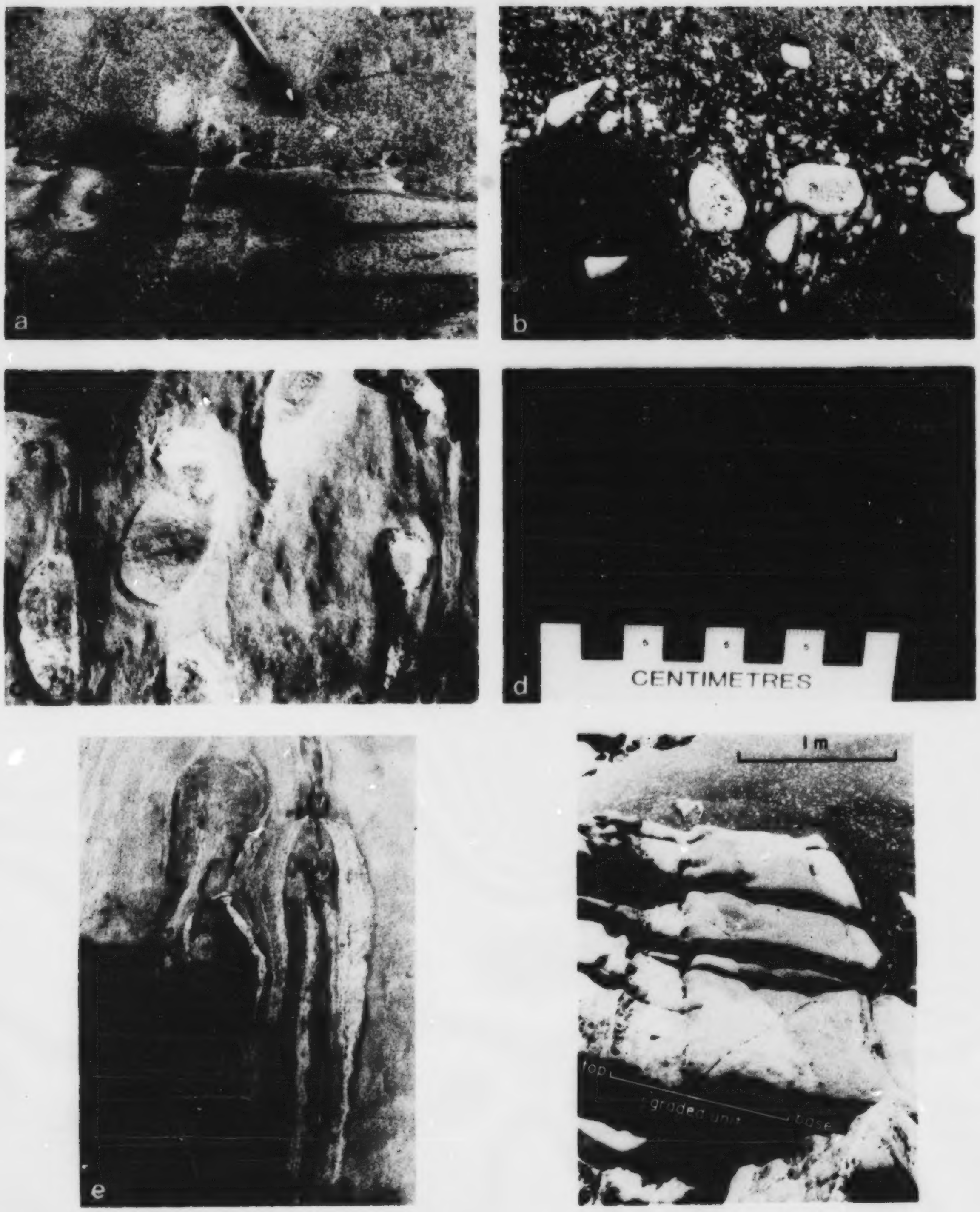


Figure 10.3. Metasedimentary rocks from the Quetico Subprovince: a) bedding contact, exhibiting flame structure, between graded feldspathic wackes, 17 km east-southeast of Beardmore; b) intraformational conglomerate near base of chloritic graded wacke unit, Jean Lake; c) arenaceous balls in pelitic matrix, an example of bed disruption resulting from dewatering, 3 km northeast of Southpine Lake; d) ultramafic metasediment exhibiting cross-lamination, Jean Lake; e) folded and altered mafic layers (dikes?) within highly sheared metasedimentary rocks, Southpine Lake; f) decimetre-scale graded bedding with tops to left, in chloritic wackes, Jean Lake.

Within the Quetico Subprovince, relatively low-grade metasedimentary rocks may be traced into more highly metamorphosed equivalents and derived migmatites in the central and southern migmatitic sections of the subprovince; these rocks may also be traced into metasedimentary granulites at the southern boundary in the Manitouwadge area (*see* Figure 10.2b). Whereas, relatively unmetamorphosed rocks are interpreted to be turbiditic in origin on the basis of the vertical succession of primary structures within depositional units. The remainder of the Quetico Subprovince, perhaps as much as 80 to 90% by area, is made up of paragneisses and migmatites derived from wackes, the origin of which is still obscure (e.g., Giguere 1972). These rocks are intruded discordantly by, and have gradational concordant contacts with, leucogranitic bodies (e.g., Pye 1965; Carter 1975; Williams 1986, 1988; Kissin et al. 1986; Percival 1989) that are preceded by biotite-hornblende-magnetite-bearing granitoids. As the metamorphic grade increases within the sedimentary pile, reliable stratigraphic younging information becomes harder to find, with the result that the overall structure of the central part of the Quetico Subprovince is still poorly known (Sawyer 1983; Williams 1989). Bed thickness, as far as it can be recognized in the highly metamorphosed and deformed central parts of the Quetico Subprovince, remains constant between the decimetre- to the metre-scale. Pelitic beds, containing sillimanite, staurolite and garnet, are migmatized to a greater extent than rocks of psammitic composition. The metamorphic mineral assemblage quartz-feldspar-biotite is ubiquitous; minor mafic interbeds, of local distribution, may represent altered and deformed mafic intrusions (*see* Figure 10.3e), or tectonic slices (Williams 1989). These isolated mafic blocks, in addition to the presence of disrupted sedimentary units, may possibly represent small-scale olistostrome development.

Geochronology of detrital zircons obtained from wackes in the Quetico Subprovince indicate source terrane ages greater than 3 Ga to just less than 2.7 Ga (Percival and Sullivan 1988; Davis et al., 1988, 1989, 1990). Detrital zircon ages, petrography and geochemical evidence suggest the most likely source of these sediments is in the calc-alkalic volcanic arc-dominated granite-greenstone terranes to the north and south of the Quetico Subprovince. Sedimentological and structural work by Devaney and Williams (1989) and Williams (1989) demonstrated that in the northern part of the Quetico Subprovince sediment was derived from the north; whereas, along the southern side of the Quetico Subprovince, the most likely sediment source was from the south (Figure 10.4). This work is consistent with the interpretations of Ojakangas (1972) who, in neighbouring Minnesota at the western extremity of the Quetico Subprovince, demonstrated a volcanogenic origin for the wackes. The work of Sawyer (1986) demonstrated that the chemical composition of the sediments permitted a mixed mafic to felsic source, probably the adjacent Wawa and Wabigoon greenstone supracrustal rocks.

IRON FORMATION

Extremely rare, thin layers of quartz-magnetite rock, representing oxide-facies iron formation (e.g., *see* Figure 10.2b, Location 3), are recorded from wackes throughout the Quetico Subprovince (Percival 1989; Williams 1989); most are only centimetres in thickness and have no consistent regional magnetic effect on 1:50 000 scale aeromagnetic maps (Williams 1989). Some iron formations, now recrystallized to melanocratic layers rich in cummingtonite and garnet, probably represent original silicate iron formation facies. Lateral continuity of individual iron formation units is poorly known, but some layers appear to form discontinuous units traceable for several tens, if not hundreds, of kilometres in the north-central part of the Quetico Subprovince (Williams 1989).

CONGLOMERATE

Polymictic, volcanically derived, matrix- and clast-supported conglomerate units, several metres thick, occur only sporadically and with no preferred distribution within otherwise rather monotonous wackes. One area where conglomerate is exposed occurs north of Terrace Bay, at Southpine Lake, near the southern margin of the Quetico Subprovince (*see* Figure 10.2b, Location 4). Lateral and vertical limits of each conglomerate occurrence are poorly known, but the occurrences are apparently restricted to no more than a few isolated outcrops. The conglomerates are enclosed in normal wacke; it is inferred, therefore, that the conglomerates are resedimented, volcanic-derived detritus. Support for this comes from the southern part of the Quetico Subprovince, where a band of fine-grained felsic volcanic breccia and derived conglomerate have been mapped close to the contact with the Wawa Subprovince from Big Duck Lake (*see* Figure 10.2b, Location 5) to Killala Lake (*see* Figure 10.2b, Location 6; Coates 1970; Carter 1988; Williams 1987b, 1989). If the hundred or more kilometres distance from a preserved volcanic source to the north are considered, the inference that these proximal-facies volcanic rocks were derived from detritus that had been transported northwards from the Wawa volcanic centres, now only 20 km to the south appears to be more reasonable.

ULTRAMAFIC WACKES, SILTSTONES AND BRECCIAS

Rocks of ultramafic and related compositions, of uncertain origin, occur sporadically throughout the Quetico Subprovince (Williams 1987b, 1988, 1989). These rocks are not large enough or sufficiently rich in ferromagnetic minerals to cause a noticeable anomaly on high resolution aeromagnetic maps.

Ultramafic and related chloritic sedimentary rocks have been found on the shores of, and islands within, Jean Lake (*see* Figure 10.2b, Location 7), about 25 km south of Beardmore (Pye 1965). For 1 km across strike, and several kilometres along strike, steeply dipping, southerly younging

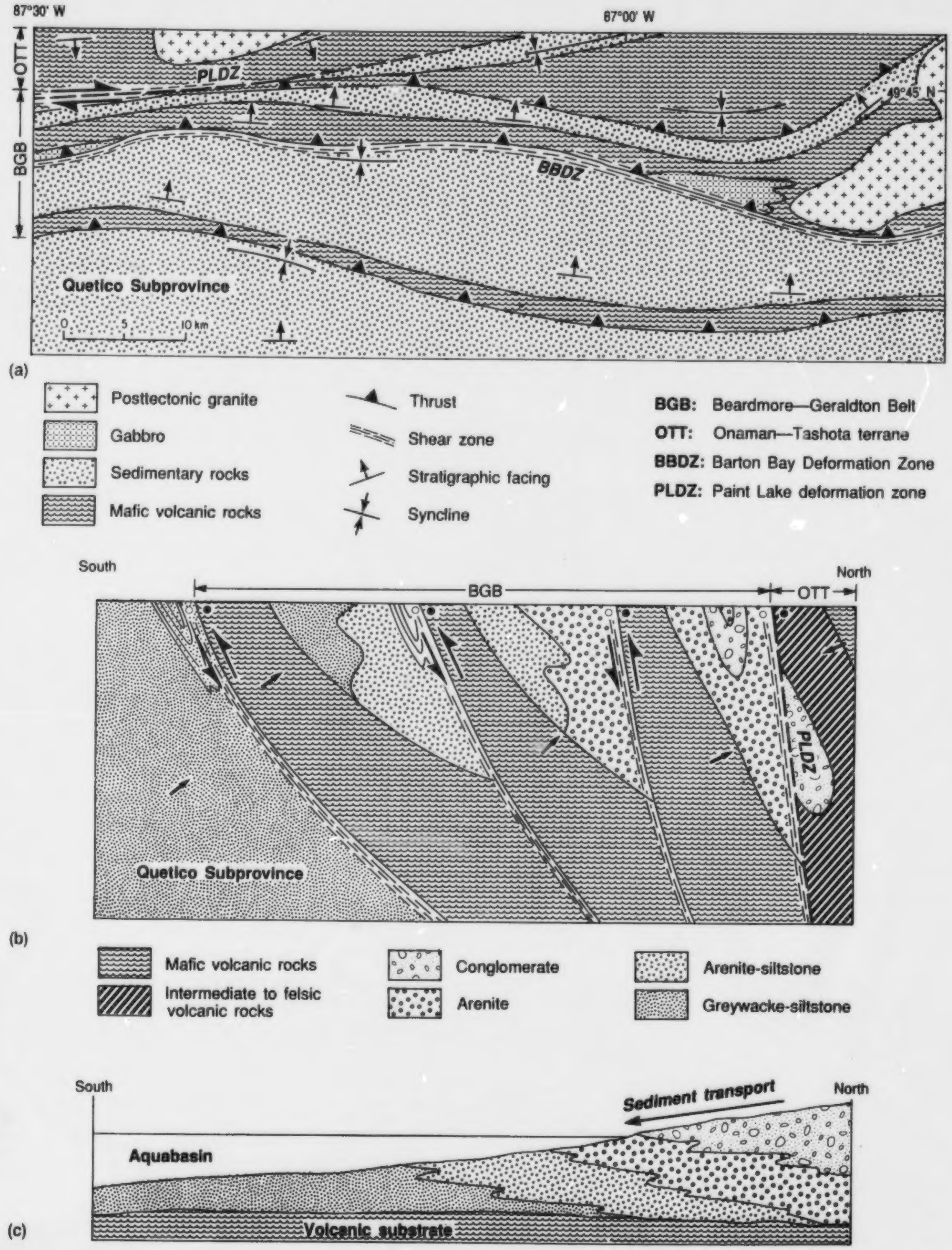

87°30' W
87°00' W
OTT
BGB
PLDZ
BBDZ
49°45' N
Quetico Subprovince
0
5
10 km
(a)
Posttectonic granite
Gabbro
Sedimentary rocks
Mafic volcanic rocks
Thrust
Shear zone
Stratigraphic facing
Syncline
BGB: Beardmore—Geraldton Belt
OTT: Onaman—Tashota terrane
BBDZ: Barton Bay Deformation Zone
PLDZ: Paint Lake deformation zone
South
North
BGB
OTT
PLDZ
Quetico Subprovince
(b)
Mafic volcanic rocks
Intermediate to felsic volcanic rocks
Conglomerate
Arenite
Arenite-siltstone
Greywacke-siltstone
South
North
Sediment transport
Aquabasin
Volcanic substrate
(c)

Figure 10.4. Geological map and cross sections of the Beardmore–Geraldton Belt (*modified from* Devaney and Williams 1989; Williams 1990 and references therein): a) geological sketch map of the Beardmore–Geraldton area, showing the generally east-striking, north-facing units of metavolcanic and metasedimentary rock that form the northern border to the Quetico Subprovince ; b) north-south oriented idealized cross section, representing about 20 km across the Beardmore–Geraldton Belt, no vertical exaggeration;c) prethrusting and rotation palinspastic reconstruction of the rocks within the Beardmore–Geraldton Belt.

wackes take on a pale to dark green colour caused by the increase in proportion, of up to 90%, of metamorphic chlorite and amphibole. Feldspathic, graded wackes pass upwards into pale to medium green chloritic (modal chlorite and amphibole 50%) wackes containing climbing ripples, scours, and cross-lamination (*see* Figure 10.3d), and succeeded by massive to graded chloritic wackes bedded on a decimetre-scale (*see* Figure 10.3f). These rocks are overlain by several tens of metres of ungraded, typically massive, poorly stratified, chlorite-rich (modal chlorite and amphibole 70 to 90%) sedimentary rocks and breccias and contain clasts of fine-grained, pale quartz arenite, or possibly recrystallized chert fragments. Vein-style alteration of the most chloritic rocks gives rise to a centimetre-scale network of quartz-carbonate veins. Within the chloritic ultramafic rocks, bedding is tightly folded, atypical of the feldspathic wackes in the immediately surrounding rocks. Wackes immediately adjacent to the ultramafic rock contrast with wackes seen elsewhere in the region; these rocks contain abundant evidence of current reworking during sedimentation, such as large ripples, discordances and cross-stratification. Breccias, associated with the sediments are composed of clasts of ultramafic and quartzofeldspathic rock set in a schistose ultramafic matrix. Podiform and discontinuous layers of ultramafic and associated dark-coloured sediment occur along a strike length of 5 km with a maximum thickness of 100 m. The finely banded to massive ultramafic sedimentary rocks contain local cross-stratification and compositional grading; some of the sedimentary rocks are clearly clastic in nature, containing siliceous to ultramafic fragments set in a chloritic matrix.

In quartzofeldspathic wackes, stratigraphically overlying the zone containing the ultramafic sedimentary rocks, is a 1 m thick mafic to ultramafic layer that lacks stratification, grading or imbrication. This layer is composed predominantly of actinolite and contains fresh, round boulders of gabbro and angular fragments of quartz-mica schist.

The chlorite- and amphibole-rich sedimentary rocks probably represent detritus derived from nearby ultramafic bodies similar to those described from the Franciscan Complex (Lockwood 1971; Carlson 1984; Robertson 1990), and from the Miura Peninsula (Kano and Masuda 1975). These rocks may have developed from the erosion of ultramafic protrusions and masses described in the following section. The bouldery layer and inclusion-rich rocks may be sedimentary mass flow deposits (Williams 1989).

SUMMARY

The sedimentary environment, represented by the rather monotonous wackes, is a submarine basin of great lateral extent, filled with rocks that exhibit little compositional variation and have sedimentary features typical of unstructured turbidite fans (P.W. Fralick, Lakehead University, personal communication). Sedimentation was rapid and reworking was minimal. Detritus provenance was probably from the north, in the northern and central parts of the basin, and may have been from the south, in the southern parts of the basin. There is as yet no unequivocal evidence on this point; indeed, sediment transport may have been along the length of the basin. There is no evidence of basement to the deposits. Along the northern boundary of the western part of the Quetico Subprovince, the timing of sediment deposition is constrained by the youngest zircon within the sediment, 2.698 Ga, and the oldest intrusion at 2.688 Ga (Percival and Sullivan 1988). Recent study of Quetico Subprovince sedimentary rocks indicates that the lithofacies assemblages have significant implications on our understanding of the late Archean development of the Superior Province (*see* Tectonic Interpretation and Summary). Whatever model is proposed to explain the development of the basin must also account for the presence of ultramafic rocks, the lack of mafic volcanic rocks and a volume of sediment deposited within a period of a little more than 10 million years, hitherto, unrepresented within the Superior Province.

Igneous Rocks

Metamorphosed rare mafic and felsic extrusive rocks, abundant felsic to intermediate intrusions and an uncommon, but widely scattered suite of problematic gabbroic and ultramafic rocks, make up the igneous rocks within the Quetico Subprovince. All the plutonic gneissic rocks but the latest discordant undeformed examples of both leucogranite and peraluminous granite are metamorphosed. On the basis of geochronological, geochemical and inclusion evidence (e.g., Day and Weiblen 1986; Percival 1989), early I-type (White and Chappell 1983) intrusions consist of hornblendites, diorites, syenites and tonalites, and contain mafic and ultramafic xenoliths that are dominated by amphibole. These typically foliated intrusions are cut by, or seen as inclusions within, leucogranite plutons, some of which contain abundant paragneiss xenoliths and screens. The mafic component of this later, often unfoliated intrusive suite is dominated by biotite and/or muscovite. These granites, consist of both one- and two-mica granites and are of mixed origins. Some of the granites are clearly peraluminous, probably derived from a metasedimentary source (Smith and Williams 1980; Day and Weiblen 1986; Percival 1989; Williams 1989) and are designated "S-type" (White and Chappell 1983); whereas, others may be derived by melting of a crustal igneous source, or perhaps a sedimentary source containing much mafic and tonalitic material. The reconnaissance nature of the mapping within the Quetico Subprovince, and the small number of intrusions that have been examined petrographically and geochemically, is such that confident distinction between granites of sedimentary or igneous origin is probably unjustifiable.

Figure 10.5 depicts chemical data for tonalites, biotite leucogranites, and peraluminous granites and shows the chemical overlap between the 3 suites of igneous intrusions. The 2 main groupings are: a granitic group, in which distribution of leucogranites of lower crustal derivation, and S-type granites are coincident (Smith and Williams 1980; Kissin and Zayachkivsky 1985; Williams 1989); and a tonalite group, in which early, mesocratic tonalites (Williams 1989; Day and Weiblen 1986) cluster with later, sheeted leucocratic tonalites (Williams 1989; Kissin and Zayachkivsky 1985). Granitic rocks show moderate to strong europium depletion, and low concentrations of heavy rare earth elements. Tonalitic rocks have flatter patterns and less, if any, europium anomaly. These data suggest that the granitoids may represent magmas that have fractionated plagioclase feldspar and left garnet in their source residuum, whereas the tonalites exhibit little evidence of plagioclase fractionation.

Swarms of thin mafic sheets, probably dikes, cut deformed wackes near the southern margin of the Quetico Subprovince.

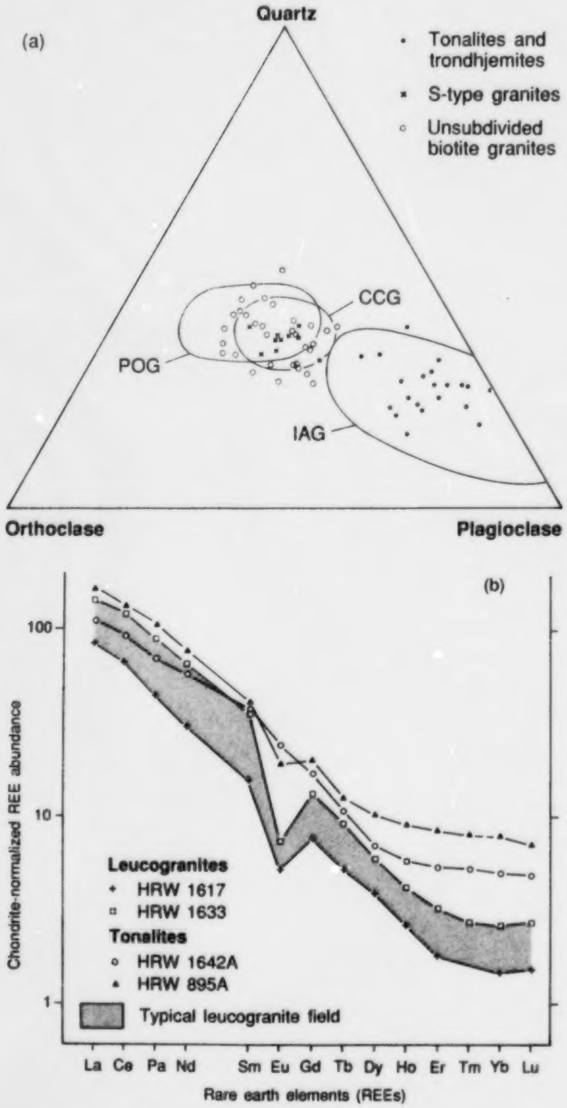


Figure 10.5. Geochemical data of tonalite, and combined leucogranitoid and peraluminous granites from the Quetico Subprovince: a) CIPW Normative quartz-orthoclase-plagioclase(QOP) triangular plot. Also plotted, *modified from* Maniar and Piccoli (1989), are the fields for post-orogenic (POG), continental collision (CCG) and island arc (IAG) granitoids; b) Chondrite-normalized rare earth element plot of 2 selected samples of tonalite, and 2 of biotite leucogranite from Williams (1989).

VOLCANIC ROCKS

Mafic volcanic rocks are virtually absent from the Quetico Subprovince, this is partly a function of how the subprovince has been defined (Card and Ciesielski 1986). A few, poorly known mafic volcanic rocks occur in the eastern part of the Quetico Subprovince in Langemarck Township (Giblin 1968). Mafic volcanic rocks occur just outside the northern margin of the Quetico Subprovince, especially within the Beardmore–Geraldton Belt; these are described in the chapter on the Wabigoon Subprovince (*see* Figure 10.4).

Felsic volcanic rocks occur along the south margin of the Quetico Subprovince, as previously mentioned under "Conglomerate" (*see* Figure 10.2b, Location 4). Clasts are composed of pale buff-coloured, quartz- and feldspar-phyric volcanic rock, up to several decimetres in diameter, set in a psammitic to pelitic matrix (Williams 1987b, 1989). Volcanic rocks in the southern part of the Quetico Subprovince are probably derived from volcanic sources to the south within the Wawa Subprovince (Williams 1988).

ULTRAMAFIC INTRUSIONS AND OTHER BODIES

Plutons, pods and layers of ultramafic composition occur within wackes of the Quetico Subprovince at over a score of localities, as intrusions (Irvine 1963; Watkinson and Irvine 1964), as armoured relics intruded by migmatitic veins, as lenses concordant with lithologic layering (Lehto 1975; Williams 1987b, 1988, 1989) and as enclaves within granitic rocks (Williams 1987b, 1989). These masses, commonly a few metres across, are rarely demonstrably intrusive; their contacts are now the site of metasomatic exchange; their original contact relationships with their host rocks are obscured by buffer zones consisting of fibrous and platy chlorite, actinolite and biotite as at Helen Lake (*see* Figure 10.2b, Location 8). Close to the Quetico Fault in the Crooked Pine Lake area, near Atikokan (Pirie 1978), hornblendite and peridotite intrusions grade into more feldspathic varieties such as gabbro (*see* Figure 10.2, Location 9). Watkinson and Irvine (1964) suspected that the bodies were formed by crystallization of primary olivine, pyroxene and amphibole from a mafic or ultramafic magma which was intruded into deeply buried metasedimentary rocks. Serpentinization occurred subsequently, during regional metamorphism. In contrast, evidence of fine-grained, chloritic sedimentary rocks, flows and breccias suggests that easily erodible buoyant, diapiric protrusions of serpentinous ultramafic rock were available for erosion (Williams 1989).

GNEISSIC AND FOLIATED TONALITE SUITES

Tonalite and diorite form generally foliated, steeply dipping, concordant sheets, usually less than 1 km in thickness, which are intrusive into paragneisses and metasedimentary migmatites in the central and southern parts of the Quetico Subprovince (Day and Weiblen 1986; Leclair and Nagerl 1988; Percival and Stern 1984; Percival 1989; Kissin et al. 1986; Kissin and Zayachkivsky 1985; Kissin and Archibald

1988). Coates (1972) reported tonalites from the Black Sturgeon Lake (*see* Figure 10.2b, Location 10) area at the Wabigoon Subprovince boundary. Zircons in these rocks have been dated at 2688 and 2687 Ma (Percival and Sullivan 1988; Davis et al. 1989, 1990). The rocks may be foliated or unfoliated, depending on whether they are intrusive into a high-grade or low-grade metamorphic host respectively. The chemistry of these rocks is distinct from the truly granitic rocks which cut them (*see* Figure 10.5). Leucotonalites, recorded from the Georgia Lake area, south of Beardmore (*see* Figure 10.2b, Location 11), by Kissin and Zayachkivsky (1985) and Williams (1989), are represented by unfoliated concordant sheets and minor plutons in relatively low-grade metasedimentary rocks.

MASSIVE GRANODIORITE TO GRANITE SUITE

Spatially associated with high-grade metasedimentary rocks is a pink, magnetite-bearing biotite leucogranite that forms migmatitic veins and elongate plutons such as the Lac La Croix Batholith (*see* Figure 10.2a, Location 12), part of the Vermilion granitic complex, at the Minnesota–Ontario border (Southwick and Sims 1980; Day and Weiblen 1986; Bauer 1986). Elsewhere, potassium-feldspar porphyritic granite and biotite leucogranite make up concordant, elliptical and irregular plutons that are ubiquitous within the Quetico Subprovince (e.g., Carter 1975; Coates 1970; Williams 1989) especially in Quetico Provincial Park (Percival and Stern 1984). Both types are true granites in composition, (*see* Figure 10.5) contain paragneiss and mafic inclusions and may cut, or be cut by, phases of the peraluminous, muscovite-bearing leucogranite. Their origin is obscure but they may both have a tonalitic, lower crustal source, rather than a sedimentary source (Day and Weiblen 1986, Percival 1989).

MUSCOVITE-BEARING GRANITIC ROCKS (PERALUMINOUS)

White to grey muscovite leucogranite is the most abundant igneous rock type in the Quetico Subprovince. An example of such a rock type is the 200 km long by 30 km wide Sturgeon Lake batholith (*see* Figure 10.2a, Location 13; Percival et al. 1985) representing a scale of intrusion rare in the eastern and central parts of the Quetico Subprovince. Peraluminous intrusions contain much xenolithic paragneiss, have interfingering, irregular contacts with the metasedimentary host and commonly occur as leucosome in migmatite (Carter 1975, 1983). Aluminous minerals such as cordierite, garnet, sillimanite and primary and secondary muscovite are found in these bodies; accessory minerals such as tourmaline, apatite and beryl are typical of this unit as are associated rare-metal-containing pegmatitic phases, found at Georgia Lake (*see* Figure 10.2b, Location 11; Kissin et al. 1986; Kissin and Zayachkivsky 1985; Kissin and Archibald 1988; Zayachkivsky 1985). Zircons, sphene and monazite in these bodies have been dated at 2670 to 2653 Ma; zircon populations indicate a dominant isotopic component inherited from its sedimentary source (Percival and Sullivan 1988). Geochemical data (*see* Figure 10.5) from Smith and Williams (1980), Day and Weiblen 1986, Kissin and Zayachkivsky (1985), Sawyer (1987), Sawyer and Robin (1986) and Sawyer and Barnes (1988) indicate that the white leucogranites are consistent with derivation by partial melting of wackes, at a depth not much greater than that presently exposed in the central part of the Quetico Subprovince.

DIORITE AND NEPHELINE SYENITE SUITE

Rocks of this unit are found only in the extreme western part of the Quetico Subprovince, within and to the west of Quetico Provincial Park (Percival et al. 1985). The most studied of these is the Poohbah Lake Complex (*see* Figure 10.2a, Location 14; Sage 1988), which is a posttectonic, postmetamorphic intrusion, consisting predominantly of syenite and nepheline syenite that cuts, and is cut by, minor granitoid sheets of leucogranite. Only 20% of the intrusion can be accurately described as undersaturated. It has been dated as 2667±5 Ma by Kwon (1986), and is temporally equivalent to leucogranites in the Quetico Subprovince.

MAFIC SHEETS

Swarms of thin mafic sheets, up to 1 m thick, are intruded nearly concordantly with bedding and S_2 foliation at the southern margin of the Quetico Subprovince, closely adjacent to the Wawa Subprovince, north of Big Duck Lake (*see* Figure 10.2b, Location 5) and in the Kashabowie area (Sawyer 1984). Some of the mafic sheets are composite in nature and contain felsic central parts, exhibiting hybridization and pillow structures (Williams 1989). These sheets become less common northwards, cut early fabrics, but are often folded or boudinaged on a decimetre-scale and contain a schistosity parallel with that of their host (*see* Figure 10.3e). These sheets are associated with, and cut tonalites intrusive into, metasedimentary rocks in the Manitouwadge area. Regionally, they are cut by leucogranite sheets and migmatitic veins and are therefore considered to be early syntectonic in age.

MAGMATIC INTERPRETATION

Timing

Isotopic ages agree with the field-based scheme for the relative timing of intrusion of the igneous rocks. The gneissic and foliated tonalite suites followed the accumulation of the sediment pile by only 5 to 10 million years, while the leucogranites and peralkalic to alkalic rocks intruded some 10 to 20 million years later (Figure 10.6). The relationship of intrusion to deformation is less clear because of the differing tectonic levels exposed within the Quetico Subprovince; the response of the high-grade, ductile metamorphic core of the subprovince contrasts with that of the relatively low-grade flanks. For this reason, the late-stage S-type granitoids appear to be posttectonic in the external regions, but are moderately to highly deformed in the high-grade regions.

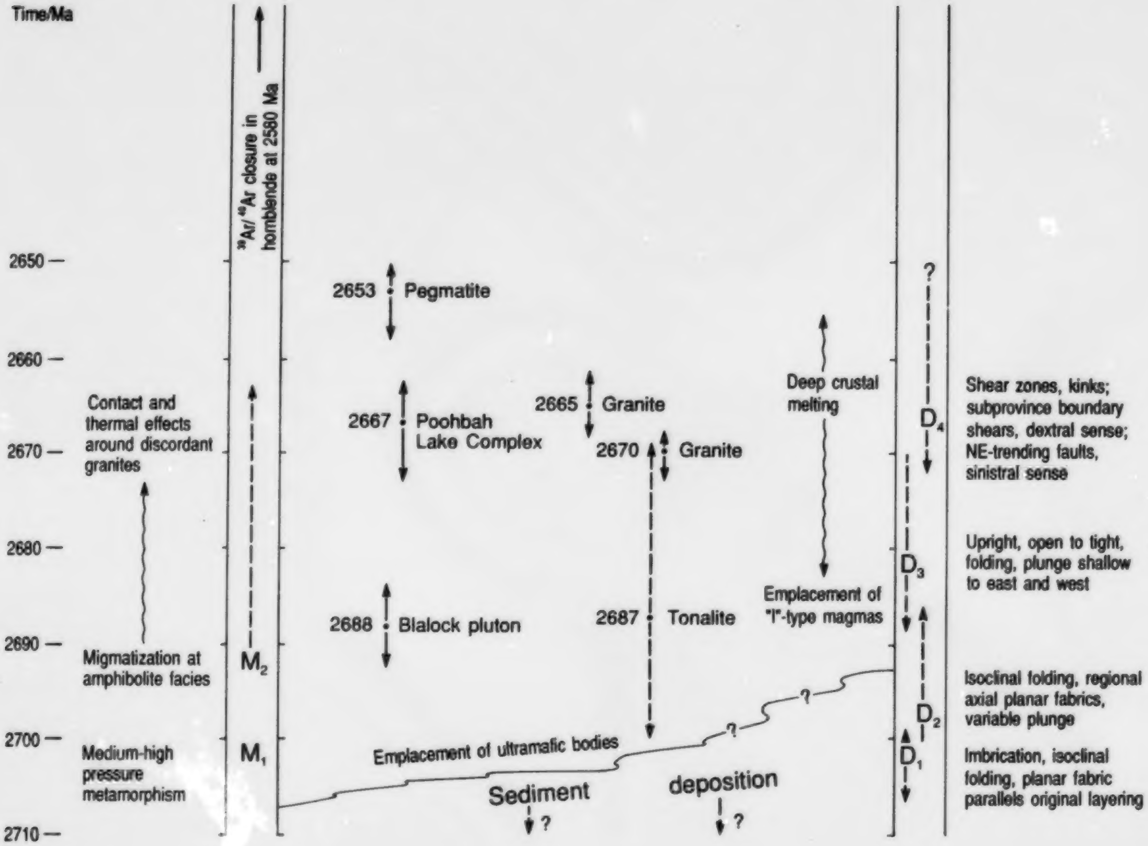

Time/Ma
2650
2660
2670
2680
2690
2700
2710
39Ar/40Ar closure in hornblende at 2580 Ma
Contact and thermal effects around discordant granites
Migmatization at amphibolite facies
Medium-high pressure metamorphism
M2
M1
2653 Pegmatite
2667 Poohbah Lake Complex
2665 Granite
2670 Granite
2688 Blalock pluton
2687 Tonalite
Deep crustal melting
Emplacement of "I"-type magmas
Emplacement of ultramafic bodies
Sediment
deposition
?
D4
D3
D2
D1
Shear zones, kinks; subprovince boundary shears, dextral sense; NE-trending faults, sinistral sense
Upright, open to tight, folding, plunge shallow to east and west
Isoclinal folding, regional axial planar fabrics, variable plunge
Imbrication, isoclinal folding, planar fabric parallels original layering

Figure 10.6. Diagram indicating chronologic sequence of events, such as the timing of sedimentation, emplacement of intrusions, deformation (D) and metamorphism (M) (Sources referenced in text).

Depth of Emplacement

The clearly intrusive ultramafic bodies were emplaced at least as deep as 8 km (equivalent to 0.25 GPa), since amphibole was on the liquidus at the time of their crystallization (Watkinson and Irvine 1964). There are insufficient data to confirm whether these bodies represent liquids or cumulates. It is unlikely that ultramafic magma intruded as a liquid into a wet sedimentary pile. Some of the bodies in the Quetico Subprovince may have been cumulates derived from mafic magma, or protrusions, emplaced diapirically into the sedimentary pile as buoyant serpentinite. The smallest bodies, 1 to 5 m in diameter, are unlikely to be diapirs and were perhaps emplaced tectonically. Similar occurrences of ultramafic rocks have been observed in the Franciscan Complex of California by Lockwood (1971) and Carlson (1984), and in southwest Newfoundland (Fox and van Berkel 1988).

The gneissic and foliated tonalite suite of intrusions, invariably metamorphosed in the central part of the subprovince, were affected by early deformation structures. These intrusions were emplaced within the sedimentary pile during tectonism because they cut early gneiss fabrics. Elsewhere, especially along the northern margin of the subprovince, tonalitic sheets are almost unfoliated (Kissin et al. 1986).

The leucogranites are syntectonic to posttectonic in character, depending on position within the subprovince. Bodies close to the outer parts of the subprovince tend to show less deformation than those close to the centre. If produced by *in situ* primary magmatic crystallization, the presence of muscovite would indicate that the magma may have been intruded at pressures of greater than 0.4 GPa (Hyndman 1981). Some muscovite in these rocks is clearly of replacement origin; this reflects metamorphic breakdown of feldspar during the high-temperature M_2 phase (Tabor et al. 1989). Leucogranites are less common in the relatively low grade, northern parts of the Quetico Subprovince, but are characteristic of the high-grade migmatitic core (Percival and Williams 1989). Tectonic thickening processes may have been constrained in these marginal zones such that partial melting of sediment was insufficient to allow rise of magma to the present erosion surface.

Source Materials

Evolution of the intrusive igneous rocks, with time, is most succinctly described as a change from mantle-derived rocks to crustal-derived rocks. This aspect of petrogenesis was first alluded to by Smith and Williams (1980), but was confirmed by Sawyer (1987), Kissin and Archibald (1988) and Percival (1989).

Early granitoids are probably derived from a combination of mantle and lower crustal sources, (I-type), based on their calc-alkalic chemistry (e.g., Smith and Williams 1980), rare earth element patterns (*see* Figure 10.5), varied mafic and ultramafic xenoliths and zircon lead signature (Percival and Sullivan 1988). Later granitoids, specifically the two-mica leucogranites, are S-type (Percival 1989) on the basis of their major, trace and rare earth element chemistry (Sawyer 1987; Kissin and Archibald 1988), inherited zircon populations (Davis et al. 1989, 1990; Percival and Sullivan 1988) and spatial relationships with highly metamorphosed sedimentary rocks (*see* Figure 10.5).

Derivation of the relatively late stage undersaturated to saturated syenites and sporadic lamprophyres has been associated with subduction related processes by Wyman and Kerrich (1988).

STRUCTURAL DEVELOPMENT

The Quetico Subprovince responded, perhaps diachronously, over a period of 40 million years (*see* Figure 10.6) between approximately 2700 to 2660 million years, to tectonic stresses in a variety of ways. Early deformation featured localized soft-sediment deformation, slumping and recumbent folding (D_1) which was progressively followed by layer-parallel shearing and associated folding and regional-scale fabric formation (D_2). Both bedding and the new planar anisotropy (S_2) were then subjected to upright folding (D_3) and localized shearing (D_4). The extent of structural analysis and the exposure of bedrock is variable throughout the Quetico Subprovince. This variability is such that recognition of all the phases is restricted to the few areas where detailed work has been carried out. There is general agreement over the style, processes and relative timing of deformation, which are depicted in Figure 10.7. Despite all the deformation and metamorphic recrystallization, most descriptions of metasedimentary rocks from the Quetico Subprovince make mention of the fact that the predominant stratigraphic-facing direction, as determined by sedimentological means, is north (e.g., Carter 1984, 1987, 1988; Harris 1970; Perdue 1938; Williams 1988).

The following sections detail the relative timing of deformation events, their characteristic styles and their geometries.

D_1 Deformation

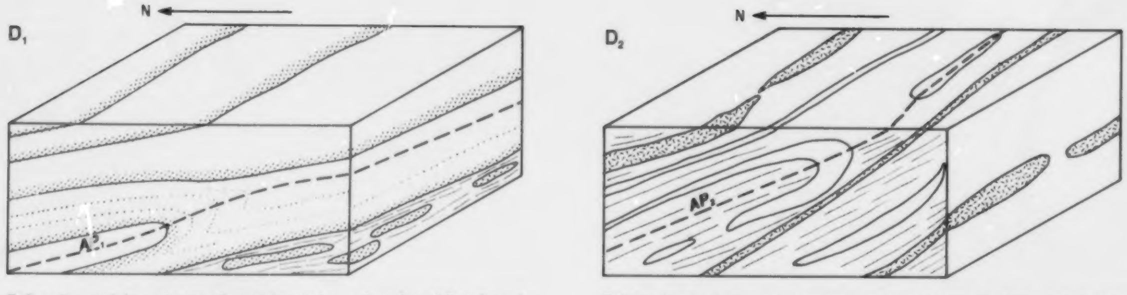
No documented examples of prefabric slump folds have been reported from within the Quetico Subprovince, but adjacent sedimentary rocks, which may be considered transitional into the Quetico, do contain these structures. Slump and soft-sediment deformation have been documented by Hudleston (1976) in sedimentary rocks of the Vermilion greenstone belt in Minnesota, southwest of the Quetico Subprovince. Similarly, slump structures have been reported from the Beardmore–Geraldton Belt, which is structurally transitional into the northern part of the Quetico Subprovince (*see* Figure 10.4).

Sawyer (1983), at Kashabowie (*see* Figure 10.2a, Location 1), considered that early folds formed in a recumbent orientation, and were northwestward facing and vergent. Bauer (1985), in Minnesota, demonstrated the need for recumbent D_1 structures in order to explain the downward-facing rocks on later D_2 folds. In Ontario, the variability of structural-facing orientations of early fold structures (Borradaile 1982; Kehlenbeck 1976, 1984; Lehto 1975), of tight to isoclinal steeply plunging folds, or periclinal, variably plunging folds transected by a cleavage, suggest that some deformation occurred prior to imposition of the regional D_2 planar fabric. Borradaile (1982), who determined that these early folds had an axial planar cleavage, inferred that the curving hinge lines of these early folds was not due to refolding, but to inhomogeneous strain. An S_1 planar fabric, most clearly developed in high- not low-grade rocks in the western end of the Quetico Subprovince (Bauer 1985) and in much of Ontario, is seen folded around D_2 folds and reworked in their limbs to become an S_2 fabric. The general lack of unequivocal stratigraphic-facing indicators in the high-grade rocks makes determination of structural facing difficult. As a result, identification of D_1 folds and the distinction of S_1 fabric from original bedding is problematical.

D_2 Deformation

Bedding-parallel shearing deformation lasted into a phase of metamorphic recrystallization and associated folding. Complex fold cleavage–structural facing relationships (Borradaile 1982; Kehlenbeck 1976, 1984) result from the superposition of nearly coplanar D_1 and D_2 fold. The S_2 cleavage often parallels both S_1 and lithologic layering, making the discrimination between the 2 phases difficult or equivocal. For example, Stott and Schnieders (1983) were unable to recognize any D_1 folds; their D_1 and D_2 is equivalent to the D_2 and D_3 in this work.

Second phase folds characteristically have steep to variable plunges; fold axes and mineral-stretching and intersection lineations often plunge downdip of axial surfaces. These folds have a typical shallow plunge, in the Manitouwadge area and along the southern border of the Quetico Subprovince but the plunge may be due to subsequent dextral shearing along the subprovince boundary (Williams and Breaks 1989). Sawyer (1983), and Stott and Schnieders (1983) who worked to the north of the Shebandowan region at Kashabowie, where the Quetico Subprovince is atypically narrow, showed that second (main) phase folds tended to have shallow plunges. Similarly, Bauer (1985) reported shallow-plunging D_2 folds along the northern margin of the Quetico Subprovince, in Minnesota. Associated with this D_2 deformation in the high-grade, biotite-bearing metasedimentary rocks and migmatites, is a prominent cleavage in low-grade rocks and a schistosity in higher grade rocks,



Soft-sediment deformation: sediment dewatering, submarine sliding, breccia formation and quartz veining.

Strain: inhomogeneous, focussed on pelitic layers and bedding-parallel shears.

Stratigraphic facing: predominantly north.

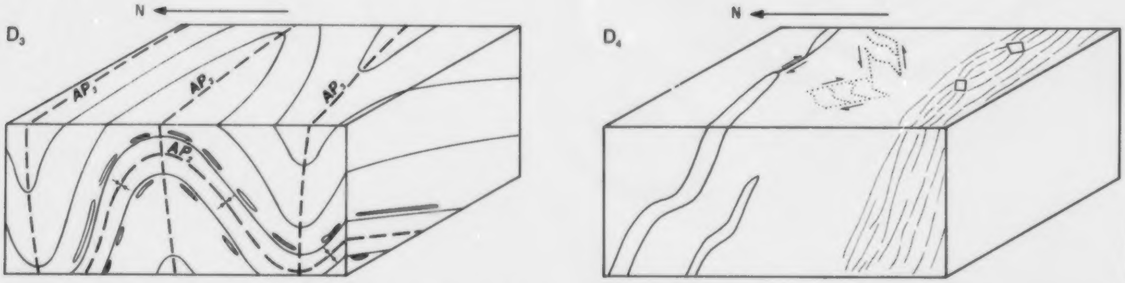
Folding: isoclinal, bedding-parallel planar fabric, axial planes (AP) bedding-parallel, variably oriented axes.

Metamorphism: medium to high pressure, emplacement of ultramafic bodies.

Folding: isoclinal, pervasive, regionally east-trending L-S fabric, axes with curvilinear plunge in schistosity. Layer-parallel shearing, progressive obliteration of bedding and way-up indicators.

Intrusion: emplacement of tonalite and leucogranite.

Metamorphism: total recrystallization of primary minerals during amphibolite-facies metamorphism. Migmatization in central and southern parts of Quetico, folding and boudinage of leucosomes.



Folding: upright, with shallow east- or west-plunging axes. Steepening of tectono-stratigraphic units.

Intrusion: continued development and emplacement of granitic rocks.

Metamorphism: continued production of differentiated layering, migmatitic leucosomes at amphibolite and granulite facies.

Boudinage of veins and intrusive sheets during regional extension along length of subprovince. Formation of sigmoidal veins and extensional shears within shear zones.

Figure 10.7. Four schematic isometric diagrams, viewed from a southwest perspective, showing characteristic styles and associated planar and linear fabrics of the main progressive stages of deformation(depiction of D_4 structures *modified from* Sawyer 1983).

composed mainly of biotite that is predominantly axial planar to these folds. In the Kashabowie area, Sawyer (1983) demonstrates that S_2 cleavage and/or schistosity transects these folds, a feature consistent with dextral transpression. In most places S_2 cleavage and/or schistosity completely overprints any earlier S_1 cleavage fabric, and in heterogeneously developed zones of shearing, obliterates evidence of bedding.

The state of strain in sedimentary rocks within the Quetico Subprovince, even within high-grade metamorphic rocks, is very heterogeneous. Highly sheared rocks occur in narrow zones, less than 1 or 2 km wide, separate much greater thicknesses of rock that exhibit well-preserved primary structures (Williams 1989).

D_3 Deformation

Upright to inclined easterly trending, shallow-plunging folds deform bedding, if preserved, and the S_2 cleavage and planar schistosity (Sawyer 1983). D_3 folds generally plunge east, but zones of westerly or variable plunge are recorded. These folds deform high-grade metamorphic mineral assemblages and migmatitic veining. Curiously, the general shallow easterly plunge of these folds would suggest that lower metamorphic grade rocks should be found in that direction. This however is not the case, allowing for the possibility that the folds developed with a shallow plunge, either before or during the metamorphic event and have not been reoriented since. This is a signature of transpression (Williams 1989).

D_4 Deformation

Small-scale shear zones cut the earlier formed planar and folded fabrics. Sawyer (1983) demonstrated that postfolding extension, rotation and shearing all tended to imply a regional compression from the south-southeast, in association with belt-parallel extension. He used late quartz veins, extension fractures and shear zones to demonstrate this tectonic regime. Similar ductile shears and semibrittle features such as kink bands are documented by Williams (1989); this indicates late-stage compression from the south-southeast.

Major Faults

Four major faults cut through the Quetico Subprovince these are the easterly trending Quetico Fault (Fumerton 1982; Bau 1979; Kennedy 1984), the Rainy Lake–Seine River fault (Fumerton 1982; Davis et al. 1989), the northeasterly trending Gravel River fault (Williams 1989), and the Kapuskasing Structural Zone (Percival 1989).

The Quetico Fault is a regional-scale dextral shear zone and fault that transects the Wabigoon–Quetico Subprovince boundary. Between Kashabowie and 45 km west of Atikokan, the Quetico Fault is the subprovince boundary. Bau (1979) calculated from the dextral offset of the Wabigoon–Quetico contact that there was a displacement of the order of 120 km. In the Rainy River region (*see* Figure 10.2a, Location 15), Poulsen (1983) described vein-type gold mineralization from the fault zone and its splays into the Wabigoon Subprovince. In the Kashabowie region, west of Thunder Bay (*see* Figure 10.2a, Location 1), the Quetico Subprovince is anomalously thin, perhaps due to strike-slip faulting along the Quetico Fault.

The Rainy Lake–Seine River fault (*see* Figure 10.2a, Quetico Subprovince boundary just south of Location 15; Fumerton 1982; Davies et al. 1989) is considered to have been an early dip-slip fault that was analogous to the dip-slip faults and shears mapped in the Beardmore–Geraldton region by Williams (1987a) (*see* Figure 10.4). The far western part of the Quetico Subprovince (Poulsen et al. 1980) is separated from a triangular part of the Wabigoon Subprovince by the Rainy Lake–Seine River fault. This part is perhaps tectonically equivalent to the Beardmore–Geraldton Belt (*see* Figures 10.2 and 10.4). The Rainy Lake–Seine River fault was subsequently transected by the Quetico Fault.

The Gravel River fault is a northeast- to east-northeast-striking shear zone and fault system displaying an oblique sinistral sense of displacement (Williams 1989). Evidence from the distribution of slightly contrasting metamorphic grades in metasedimentary rocks on opposing sides of the fault in the Dickison Lake area (*see* Figure 10.2b, Location 16; Carter 1975), suggest that the fault has a south-side-down component. The intersection of the Gravel River fault with the Quetico–Wawa subprovince boundary is hidden by Proterozoic platform deposits; the fault offsets the Quetico–Wabigoon boundary by at least 70 km, and greenstone belts of the Wabigoon Subprovince by up to 40 km (I.A. Osmani, Ontario Geological Survey, personal communication, 1990) beneath Phanerozoic rocks of the James Bay Basin. Sporadic sulphide mineralization occurs along the fault (Carter 1975).

The Kapuskasing Structural Zone, considered by Percival (1989) to be predominantly an upthrust block, locally brings deep-level, highly metamorphosed Quetico Subprovince metasedimentary rocks and associated intrusions to the surface along a narrow zone on its west side. In addition, there is evidence of considerable sinistral motion, perhaps as great as 160 km, across the Kapuskasing Structural Zone (Watson 1980; Goodings 1988), supported by regional map patterns of subprovinces, seismic reflection and aeromagnetic data (Percival et al. 1989).

North- and southeast-striking faults and fracture zones commonly cut through earlier ductile structures and can be seen most easily on recent Landsat images. Displacements across the faults are minor. One of the most prominent is the Thiel fault (*see* Figure 10.2b, Location 17) which intersects the Quetico Subprovince in the region north of the Port Coldwell alkalic complex (*see* Figure 10.2b, Location 17). The Thiel fault has an unknown displacement sense and is documented from seismic studies in Lake Superior (Cannon et al. 1989).

Significance of Structural Style and Development

There is a consensus among structural analyses of the Quetico Subprovince that early folds and associated layer-parallel shearing are coplanar, and represent the effects of progressive rotation of early structural features towards a more steeply dipping orientation by subsequent deformation. This steepening of early recumbent, thrust and fold structures is mainly inferred from detailed stratigraphic and structural work carried out at the northern boundary of the Quetico Subprovince (*see* Figure 10.4) in the Beardmore–Geraldton Belt (Devaney and Williams 1989). Note in Figure 10.4b the combined dextral and overthrust motions on volcanic-sediment contacts and the localized footwall synclines within overridden metasedimentary units. Most authors invoke a north-northwest-directed compression to explain the orientation and attitude of the folds (Bauer 1986, 1988; Bauer et al. 1989; Borradaile 1982). Several attempts have been made to relate the early, recumbent folding to that seen in adjacent greenstone belt terranes (Bauer 1985; Borradaile 1982; Kehlenbeck 1984; Tabor and Hudleston 1988; Tabor et al. 1989). In northern Minnesota, Tabor and Hudleston (1988) demonstrated at the southern boundary of the Quetico Subprovince that early shortening was followed by transpressive deformation. This involves a regional dextral strike-slip component in addition to a regional north- northwest-directed flattening deformation. This work is in agreement with that of Stott and Schnieders (1983).

METAMORPHISM

Metamorphism is considered here in some detail, because it provides constraints on the origin and history of the entire subprovince to a greater degree than is the case for most granite-greenstone subprovinces. All rocks in the Quetico Subprovince, except for some of the late-stage leucogranites and Proterozoic diabase (*see* Osmani, this volume) and alkalic intrusions (*see* Sage, this volume), were regionally metamorphosed in a regime that climaxed during the waning stages of tectonism. The following section deals with the distribution, character, causes and timing of metamorphism. A recent study (Tabor et al. 1989) has shown that the regional high-temperature metamorphism M_2 was preceded by an earlier metamorphic event, M_1, in which high-pressure mineral phases are only rarely preserved.

Distribution and Character

The gross distribution of metamorphic grades was recently reviewed by Percival (1989). Intensity of metamorphism is related to position within the subprovince such that marginal rocks tend to be of lower grade than those in the high-grade core regions. However, in the Manitouwadge region, a belt of metasedimentary granulites occurs close to, and parallel with, the southern margin of the Quetico Subprovince. These high-grade rocks may be traced northwards into amphibolite-facies migmatitic metasedimentary rocks that typify the bulk of the Quetico Subprovince.

The generally psammitic metasedimentary rocks exhibit a simple mineralogy, consisting of quartz, feldspar and biotite, which does not allow more than a broad metamorphic subdivision. Pelitic rocks offer further subdivision, using isograds determined from the appearance and disappearance of orthopyroxene, cordierite, garnet, staurolite and andalusite (e.g., Percival 1989; Tabor et al. 1989; Pirie and Mackasey 1978). High-grade metamorphic rocks, up to granulite facies, are typically found within the central and southern parts of the subprovince as lensoid masses elongate along the regional structural trend, and where tectonically uplifted to the erosion surface within the Kapuskasing Structural Zone. The lowest metamorphic grades are usually found along the northern boundary with the Wabigoon Subprovince. On the Bedrock Geology of Ontario, maps 2542, 2543 and 2545, the distribution of low-grade and high-grade metasedimentary rocks is indicated. Locally, very low-grade to greenschist-facies rocks occur along the southern boundary (Borradaile 1982), but more often, there is a rapid rise in grade north of the Wawa Subprovince and a lesser gradient in from the Wabigoon Subprovince (Pirie and Mackasey 1978). Therefore, the grade distribution is asymmetric, with the thermal and pressure maximum south of the central region, sometimes coincident with the southern margin. The metamorphic character is of high-temperature–low-pressure metamorphism (Abukuma-type), associated with abundant intrusions of granites, and the regional distribution of migmatites derived from metasedimentary rocks (Kamineni et al. 1988; Percival and McGrath 1986; Percival 1989; Williams 1989).

In the Atikokan region, at the northern margin of the Quetico Subprovince, and in neighbouring northern Minnesota, evidence of an M_1 metamorphic peak during D_1 and D_2 that produced kyanite-staurolite-biotite assemblages, supports higher pressure metamorphism prior to the introduction of the massive granodiorite to granite suite (Ayres 1978; Tabor et al. 1989).

Geothermometry and geobarometry by Percival (1989) indicate that higher temperature, migmatitic rocks of the Quetico Subprovince also represent more deeply eroded tectonic levels. This evidence suggests that tectonic thickening of the Quetico Subprovince was greater in its central and southern part. There is no evidence of folded isograds. Pressures and temperatures attained during the metamorphism of the highest grade rocks range from 0.25 GPa and 500°C, near the Minnesota border, to nearly 0.6 GPa and 780°C, at the Kapuskasing Structural Zone; typical values for the central region are in the order of 0.3 GPa and 620°C (Percival 1989).

There is little information concerning pretectonic or early syntectonic metamorphism. A study in the southern part of the Beardmore–Geraldton Belt (Williams 1989) and to the west in the Smiley–De Courcey lakes area (Kehlenbeck 1976), indicated that metamorphic grades were locally at amphibolite facies during D_2 deformation, but that facies distribution was, in part, structurally controlled by position within thrust-bounded panels (Williams 1986). There is newly acquired evidence of an early, higher pressure metamorphic event from mineral chemistries within mafic and ultramafic rocks included within the metasedimentary rocks (Tabor et al. 1989).

The main phase of regional metamorphism (M_2) occurred syntectonically and during the waning stages of deformation (Sawyer 1983) and is displayed as a thermal anticline (Percival and Williams 1989). In the central and southern parts of the subprovince, some of the highest temperature minerals, such as sillimanite, occur as oriented growths rodded along D_3 folds (Williams 1989). In the northern (Pirie and Mackasey 1978) and southern (Sawyer 1984) parts of the Quetico Subprovince, the metamorphism had reached its peak by the time of the pervasive D_2 deformation.

Causes and Timing of Metamorphism

The time relationships between deformation and metamorphism are such that tectonic thickening of the sedimentary pile and the intrusion of minor I-type granitoids, had already occurred prior to the thermal acme. Most of the large granitoid bodies display little evidence of thermal metamorphism caused by their intrusion; one exception is the relatively late body occurring within the northern part of the Quetico Subprovince (e.g., South Beatty Lake pluton, Pirie and Mackasey 1978). Steeply-dipping horizontal thermal gradients, local increase in temperature around large intrusive masses, and the general association of the highest grade rocks with abundant leucosome generation indicate that the source of heat at the present exposure level was, in part, due to burial and, in part, to upward magmatic transport. Local thermal peaks around late plutons, dated at around 2.67 to 2.65 Ga (Percival 1989), overprint regional metamorphic fabrics.

A more detailed discussion of the thermal consequences of tectonic thickening (cf., Gordon, in press) is discussed along with a detailed tectonic interpretation of the Quetico Subprovince (*see* Williams, Stott, Thurston et al., this volume).

GEOCHRONOLOGY

Early geochronological work yielded a number of Rb-Sr and K-Ar ages in the range 2.4 to 2.6 Ga, which reflects the metamorphism and subsequent cooling of the Quetico Subprovince (for a compilation of these ages, *see* Easton 1986a and 1986b). Recently determined U-Pb isotopic data on zircons, monazite and titanite (sphene) indicate ages of igneous crystallization and metamorphic recrystallization (Percival and Sullivan, 1988). Detrital zircons obtained from metasedimentary rocks have indicated that the sources of the sediments varied in age from 2698 to over 3000 Ma (Percival and Sullivan 1988; Davis et al. 1988, 1989, 1990). Figure 10.6 displays some of these recently determined ages; their locations are plotted on the Tectonic Assemblage map of Ontario that accompanies this volume (*see* maps 2575–2583 inclusive, map case). Recent unpublished Ar-Ar ages determined on hornblende and biotite indicate that cooling was protracted; rocks now exposed apparently passed through the 300°C isotherm at about 2500 Ma, some 150 million years after the metamorphic peak (D. York and P. Layer, Physics Department, University of Toronto, personal communication, 1989).

MINERALIZATION

The potential for widespread economic mineralization within the Quetico Subprovince is generally lower than that in the adjacent greenstone-dominated terranes. For this reason, little work on the local and regional controls on mineralization has been attempted. A potentially important ore deposit type comprises late-stage pegmatites that contain the rare metals lithium, beryllium, tantalum, niobium and tin.

Detailed studies of lithium prospects and rare-metal (tantalum- niobium-tin) pegmatites (e.g., Pye 1965; Breaks 1980; Kissin and Archibald 1988), especially in the Georgia Lake area (*see* Figure 10.2b, Location 11), are the only recent attempts to determine mineral potential within the Quetico Subprovince. Gold mineralization is associated with veining along the Quetico Fault (Poulsen 1983). Rare occurrences of molybdenite in the biotite leucogranites are reported by Carter (1975, 1985), in the Dickison Lake area.

Archean gold introduction may be dependent on upper mantle processes such as underplating by subcontinental lithosphere (Fyon et al. 1989). If the accretionary prism origin for the Quetico Subprovince discussed below is valid, the subprovince would have lacked both older sialic basement and subcontinental lithosphere. This might explain the rarity of gold.

TECTONIC INTERPRETATION AND SUMMARY

A brief review of the tectonics of the Quetico Subprovince follows; an analysis of the tectonics will be found in a later chapter titled "Tectonic Evolution of Ontario: Summary and Synthesis". Inasmuch as the tectonics of one subprovince is dependent on that of its neighbours, the reader is advised to refer to details of the southern part of the Wabigoon Subprovince, known as the Beardmore–Geraldton Belt (*see* Figure 10.4 and Blackburn et al., this volume). In the following paragraphs a tectonic model for the Quetico Subprovince, based upon relationships examined in the Beardmore–Geraldton area (Williams 1989), will be developed.

The Wabigoon Subprovince is composed of a linear to cuspate collage of disparate greenstone belts and abundant calc-alkalic plutonic rocks that served as a source area for volcanic and plutonic rocks for the sedimentary rocks within the Beardmore–Geraldton Belt, the southern part of the Wabigoon Subprovince (Williams 1990). The Beardmore–Geraldton Belt consists of 3 fault-bounded panels each of which contains a sedimentary upper unit unconformably overlying a volcanic lower unit. These sedimentary rocks form 3, internally southward-fining clastic units representing, from north to south, an alluvial fluvial, submarine fan and an abyssal sedimentary environment (Devaney and Williams 1989). The basaltic lower unit in each panel is identical in terms of geochemistry and primary structures; geochemistry is distinct from that of the greenstones to the north (Williams 1990). Downdip isoclinal fold axes, mineral and stretching lineations, a predominant north-facing stratigraphic sequence, south-side-down kinematic indicators on now vertical shears that parallel the lithologic layering and the apparent repetition of 3 identical mafic volcanic units have been used to interpret the repeated units as a thrust-imbricated succession (Figure 10.8; Devaney and Williams 1989). This evidence of tectonic imbrication of supracrustal sequences within the Beardmore–Geraldton Belt, marginal

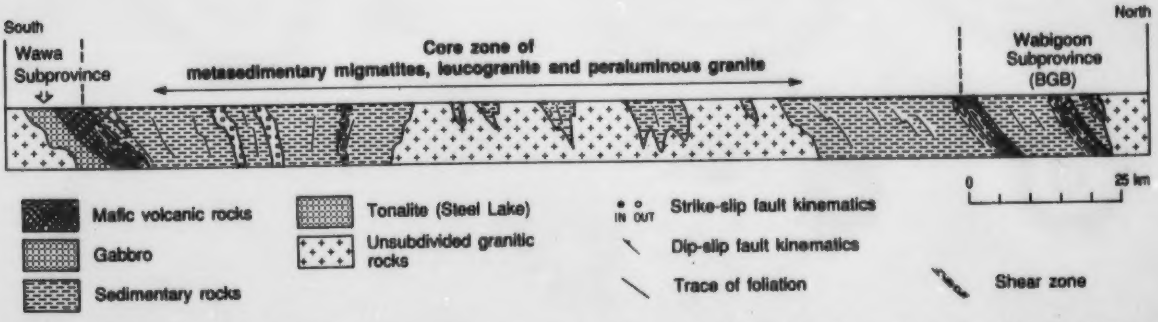

South
Wawa Subprovince
Core zone of metasedimentary migmatites, leucogranite and peraluminous granite
North
Wabigoon Subprovince (BGB)
0
25 km
Mafic volcanic rocks
Gabbro
Sedimentary rocks
Tonalite (Steel Lake)
Unsubdivided granitic rocks
IN OUT
Strike-slip fault kinematics
Dip-slip fault kinematics
Trace of foliation
Shear zone

Figure 10.8. Schematic north-trending vertical cross section across the Quetico Subprovince at long. 87°25′W. Note: 1) imbricated mafic and sedimentary units within the Beardmore–Geraldton Belt (BGB); 2) generally northward-dipping fabrics and stratigraphic units; 3) structure of granitic rocks within migmatitic core of Quetico Subprovince not known; 4) stratiform and concordant nature of tonalite sheets near southern boundary; 5) presence of ductile shear zones exhibiting both dip- and strike-slip kinematics, between imbricate units in BGB, and at the Wawa–Quetico subprovince boundary; and 6) northeastward striking Gravel River fault intersects section at low angle.

to the northern edge of the Quetico Subprovince, has allowed the palinspastic reconstruction of the stratigraphy in terms of a southerly prograding clastic wedge that has been shortened predominantly by thrusting. The major shears, originally listric thrusts, occur at the structural bases of 3 basalt-sediment couplets (Williams 1986; Devaney and Williams 1989). These thrusts and their enclosing stratigraphy were subjected to regional shortening, and became rotated to near vertical.

The current interpretation of the Beardmore–Geraldton Belt is that it represents a shortened fold and thrust belt. Interpretation of the geochemistry of the closely related Beardmore–Geraldton Belt metavolcanic rocks, and the sediment-rich character of the Quetico Subprovince suggests the subprovince can be interpreted as a fore-arc accretionary prism (Williams 1990). The lithologic (Mackasey et al. 1974; Blackburn and Mackasey 1977) and structural (Williams 1987a) transition between the Beardmore–Geraldton Belt and the Quetico Subprovince provides the key to understanding the presence and structure of the metasedimentary terrane to the south.

The Quetico Subprovince is a large linear belt of monotonous, rapidly deposited clastic sediments. It was subsequently thickened by layer-parallel shearing, during which time the predominantly northward-younging sedimentary sequences were rotated and folded into a near vertical orientation. These rocks were subjected to intense shearing, emplacement of sporadic ultramafic rocks, and local intrusion by I-type granitoids. The heating, melting and metamorphic patterns in the Quetico metasedimentary rocks may in part be the result of tectonic thickening, but a mantle-derived heat source that operated during the emplacement of the tonalites cannot yet be ruled out.

At the southern boundary of the Quetico Subprovince near Manitouwadge, granulite-facies migmatitic metasedimentary rocks are in tectonic juxtaposition with a high-grade granite-greenstone terrane of the Wawa Subprovince, along a major, northward-dipping shear zone (Williams 1989; Williams and Breaks 1989). Elsewhere, lower grade, northward-facing metasedimentary rocks are in dip-slip sheared contact with northward-facing Wawa Subprovince greenstone belt rock types (Williams 1989). These field relationships are interpreted to be the result of the Wawa Subprovince underthrusting the southern edge of the Quetico Subprovince.

After the coalescence of the 3 subprovinces, continued shortening oblique to the subprovince boundaries encouraged rotation of originally flat-lying to shallowly north-dipping thrusts and shears towards a vertical orientation. Coalescence and continued convergence also led to the development of major dextral strike-slip shear zones along the tectonic subprovince boundaries (Williams 1990).

The association of the Wabigoon and Quetico subprovinces can be interpreted in terms of a fore-arc accretionary prism tectonic model (Percival and Williams 1989; Williams 1990). The Wawa–Quetico subprovince relationship is one of a greenstone-dominated terrane underthrusting a fore-arc accretionary prism. The corresponding underthrust relationship of the Quetico Subprovince, lying structurally beneath the Wabigoon Subprovince, leads towards a universal mechanism for subprovince development by the accretion of primitive island arcs (Percival and Williams 1989; Williams 1990).

Outstanding Problems

The following problems remain to be solved through the application of field and laboratory techniques:

1. Precise characterization of the tonalitic to granitic igneous intrusions within the Quetico Subprovince must be made. Many intrusions have had their composition determined only in a reconnaissance fashion.
2. What are the controls on the metamorphic grade variation in the Quetico Subprovince? A thorough determination of pressure and temperature variation within the metamorphic belt would allow tectonic modelling to be made of this metamorphic variation.
3. What are the origins, timing, mineral potential and emplacement of the ultramafic bodies? How might these bodies relate to the origin of the Quetico Subprovince?
4. What are the relationships between the Opatica Subprovince in Ontario and Quebec and between these occurrences and the Quetico Subprovince, west of the Kapuskasing Structural Zone? Does the Opatica Subprovince in Ontario contain plutonic-dominated rocks similar to those mapped in western Quebec? If not, should the Opatica Subprovince in Ontario be renamed? Are the Opatica and Quetico subprovinces part of a continuous feature separated by the Kapuskasing Structural Zone? If not, what are the implications for the interpretation of the Kapuskasing Structural Zone?
5. A search should be made for 3-dimensional characterization of the proposed thrust fault system within the Beardmore–Geraldton Belt in order to test the accretionary prism model.

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Chapter 11

The Western Abitibi Subprovince in Ontario

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Abstract

The 2.75 to 2.67 billion-year–old "granite-greenstone"-dominated Abitibi Subprovince lies at the southern edge of the Archean Superior craton. Komatiitic, tholeiitic and calc-alkalic metavolcanic rocks formed in ensimatic oceanic settings between 2.75 and 2.70 Ga; turbidite-dominated assemblages appear to have formed between 2.70 and 2.68 Ga; and alkalic metavolcanic rocks and associated alluvial-fluvial metasedimentary rocks formed between 2.68 and 2.67 Ga. Granitoid intrusive rocks include: 2.74 to 2.69 billion-year-old tonalite-trondjhemite-granodiorite batholiths; smaller, 2.70 to 2.68 billion-year-old granodiorite intrusions; and 2.69 to 2.67 billion-year-old syenitic stocks.

The supracrustal rocks can be subdivided into: 1) komatiite-tholeiite-dominated assemblages with interflow iron formation; 2) komatiite and/or tholeiite-dominated assemblages with significant felsic metavolcanic rock (volcanogenic massive sulphide and komatiite-hosted nickel-sulphide deposits occur within this assemblage type); 3) komatiite-tholeiite-dominated assemblages without significant iron formation or felsic metavolcanic rocks (komatiite-hosted nickel-sulphide deposits occur within this assemblage type); 4) tholeiite-dominated assemblages characterized by alternating magnesium- and iron-rich units; 5) tholeiite-dominated assemblages containing thick units of either iron- or magnesium-rich units, or both; 6) ultramafic to mafic and felsic metavolcanic rocks associated with iron formation; 7) intermediate to felsic metavolcanic, fragmental rock-dominated assemblages; 8) intermediate effusive metavolcanic assemblages; 9) turbiditic metasedimentary-dominated assemblages; and 10) alluvial-fluvial metasedimentary and alkalic metavolcanic-dominated assemblages (Timiskaming-like assemblages). Most of the metavolcanic assemblages are between 2720 and 2700 Ma; however, some are 2750 to 2720 Ma and, at least locally, form stratigraphic and/or structural basement to the younger assemblages. The alluvial-fluvial metasedimentary assemblages locally rest unconformably on top of the metavolcanic-dominated assemblages and the turbidite-dominated metasedimentary assemblages. Contacts between other assemblages are either unknown or they are faults.

Tectonic fabric and mesoscopic structures are domainal in distribution and type within the supracrustal rocks of the subprovince. In general, penetrative fabric and structures parallel, and are best developed adjacent to, regional faults, large batholiths and many of the exposed assemblage boundaries. Early structures include "pre-cleavage" folds, thrust faults, and structures related to batholith emplacement. Regional shear zones and folds developed during and following batholith emplacement strike west, northwest to west-northwest, and northeast to east-northeast. Thrust faults and/or steep reverse faults are also associated with these later structures. The above structures are interpreted to have formed during protracted Neoarchean, north-south subhorizontal compression. Brittle faults related to the development of the Paleoproterozoic Cobalt Embayment and the Phanerozoic Timiskaming Rift overprint the Archean structures and strike northeast, northwest and north-northeast.

A four-stage evolutionary model is suggested for the southern Abitibi greenstone belt. 1) Formation of submarine oceanic assemblages in a region of complex micro(?)-plate interactions, perhaps caught between 2 larger converging plates located north and south of the micro-plate region. Local trenches and any large-scale trenches accommodating large-scale convergence were likely starved of sediment at this time, due to lack of emergent source areas. 2) Termination of submarine volcanism by collision of a large continental mass to the south at~2700 Ma. The collision may have been oblique, involving the 2.8 to 3.0 billion-year-old Minnesota River Valley gneiss terrane (?). 3) Tectonic thickening during collision led to emergent sediment source area(s) for post–2700 Ma turbidite deposits, including both local deposits and a massive sediment influx to develop the deposits of the Pontiac metasedimentary belt, which may represent a remnant of a major accretionary wedge. As collision continued, previously formed volcanic and turbidite deposits, including the Pontiac Subprovince, were deformed. 4) Terminal subduction, possibly involving complex plate interactions at 2685 to 2675 Ma, generated alkalic volcanic rocks and alluvial-fluvial sediments in proximity to crustal-scale shear zones. Archean lode gold mineralization generally shows a spatial relationship to these regional shear zones.

OVERVIEW OF THE ABITIBI SUBPROVINCE

The Abitibi Subprovince (Figures 11.1 and 11.2) is a Neoarchean granite-greenstone-gneiss terrane that developed between 2.8 and 2.6 Ga. The subprovince is bounded to the west by the Kapuskasing Structural Zone, a Neoarchean–Paleoproterozoic intracratonic crustal-scale overthrust (Percival and Card 1983, 1985). To the south and east, the subprovince is bounded by the Paleoproterozoic rift to passive-margin sequence of the Huronian Supergroup (Card et al. 1972; Young 1983; *see* Bennett and Dressler, this volume), the Neoarchean Pontiac Subprovince and the northeast-trending high-grade gneisses and granitoids of the

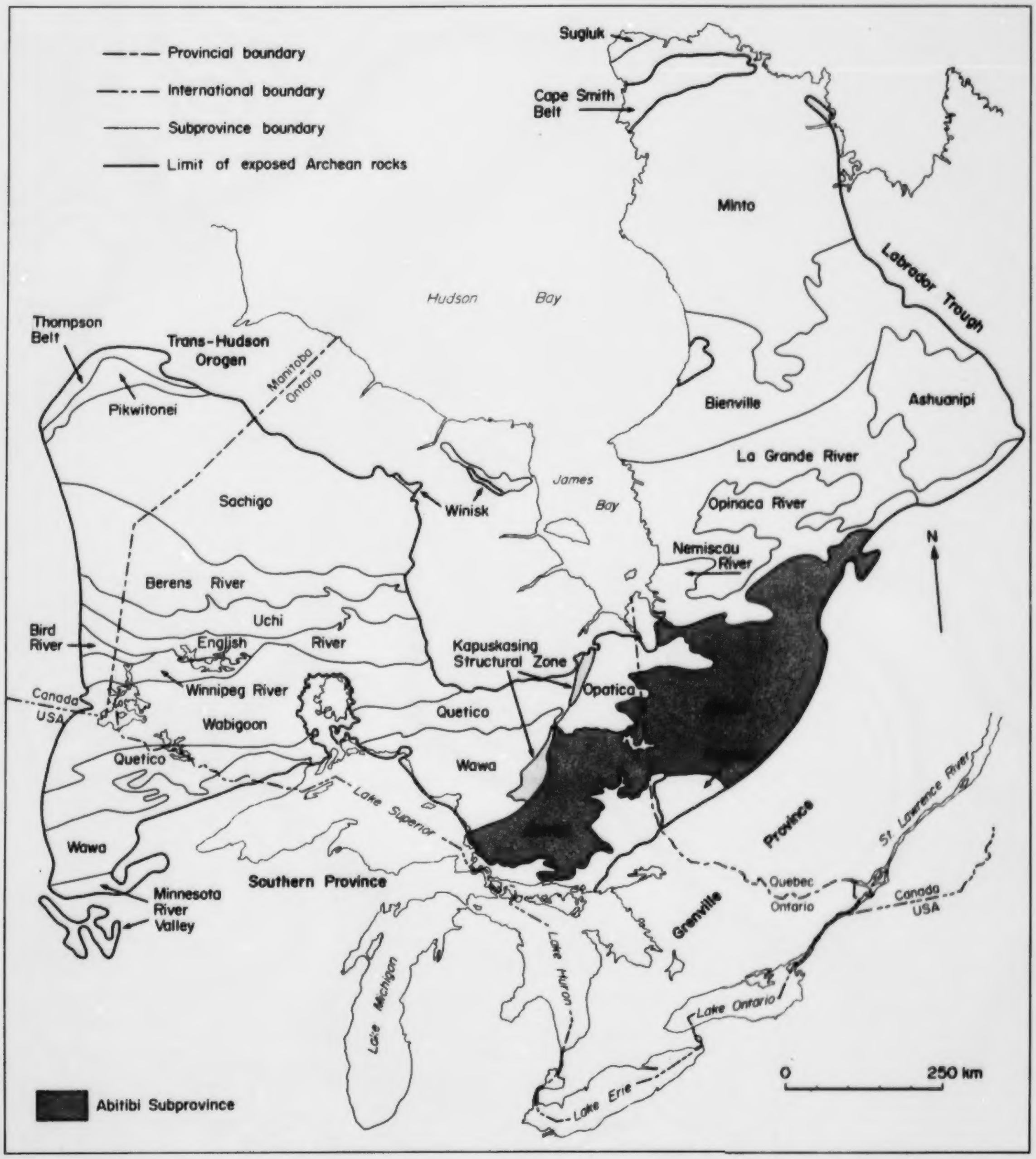
Provincial boundary
International boundary
Subprovince boundary
Limit of exposed Archean rocks
Sugluk
Cape Smith Belt
Minto
Labrador Trough
Hudson Bay
Thompson Belt
Trans-Hudson Orogen
Manitoba
Ontario
Pikwitonei
Bienville
Ashuanipi
La Grande River
James Bay
Sachigo
Winisk
Opinaca River
Nemiscau River
N
Berens River
Uchi
Bird River
English River
Kapuskasing Structural Zone
Winnipeg River
Opatica
Canada
USA
Wabigoon
Quetico
Quetico
Wawa
Lake Superior
Province
St. Lawrence River
Wawa
Southern Province
Quebec
Ontario
Canada
USA
Minnesota River Valley
Grenville
Lake Huron
Lake Michigan
Lake Ontario
0
250 km
Abitibi Subprovince
Lake Erie

Figure 11.1. Location of the Abitibi Subprovince in the central Superior Province (*modified from* Card and Ciesielski 1986).

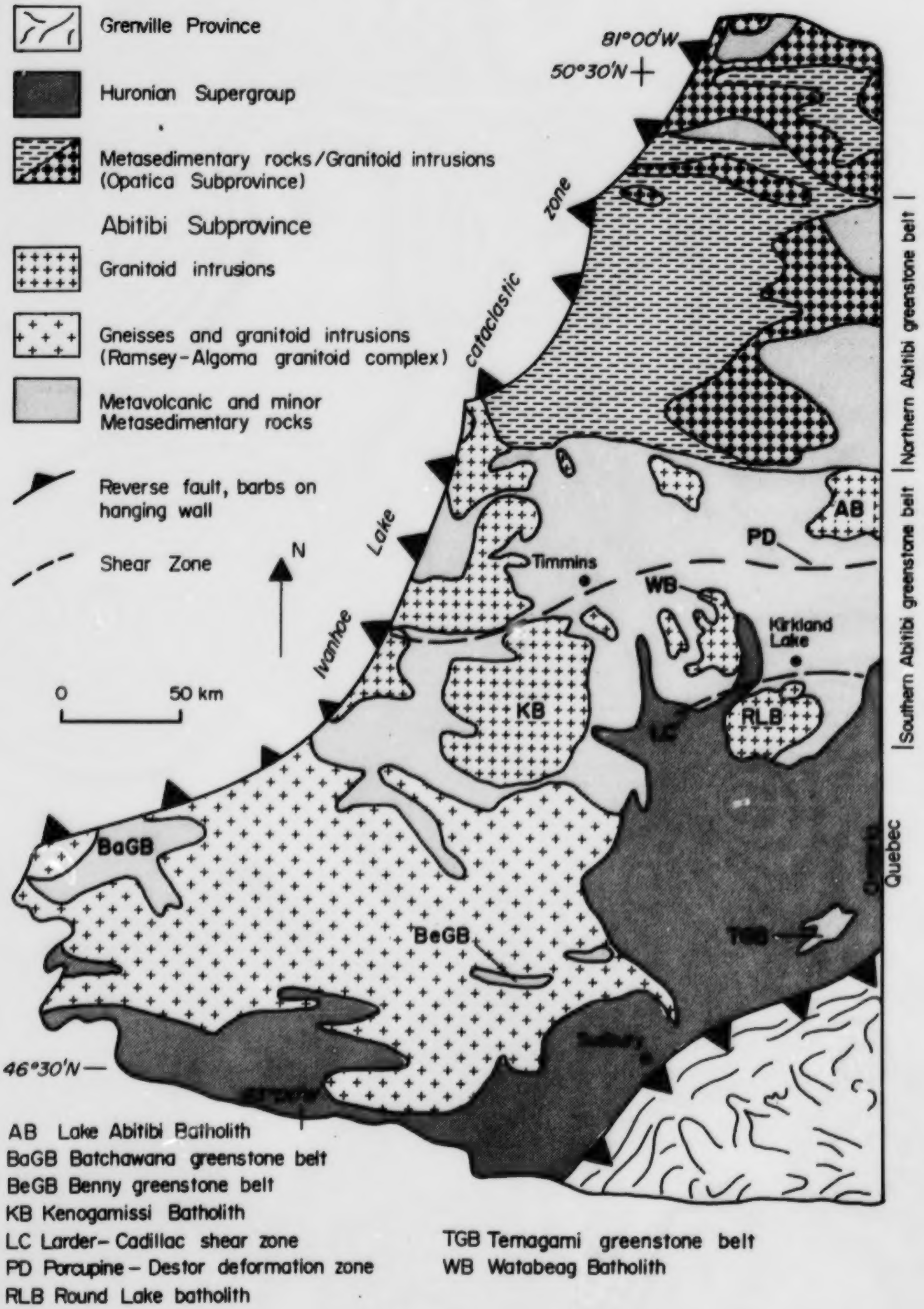
Grenville Province
Huronian Supergroup
Metasedimentary rocks/Granitoid intrusions (Opatica Subprovince)
Abitibi Subprovince
Granitoid intrusions
Gneisses and granitoid intrusions (Ramsey-Algoma granitoid complex)
Metavolcanic and minor Metasedimentary rocks
Reverse fault, barbs on hanging wall
Shear Zone
N
0
50 km
81°00'W
50°30'N
46°30'N
Ivanhoe Lake cataclastic zone
Northern Abitibi greenstone belt
Southern Abitibi greenstone belt
AB
PD
Timmins
WB
Kirkland Lake
KB
LC
RLB
BaGB
BeGB
TGB
Quebec
AB Lake Abitibi Batholith
BaGB Batchawana greenstone belt
BeGB Benny greenstone belt
KB Kenogamissi Batholith
LC Larder-Cadillac shear zone
PD Porcupine-Destor deformation zone
RLB Round Lake batholith
TGB Temagami greenstone belt
WB Watabeag Batholith

Figure 11.2. General geology of the Abitibi Subprovince in Ontario. Division between northern and southern Abitibi greenstone belt *after* Dimroth et al. (1982).

Grenville Province that were deformed during Mesoproterozoic northwest-directed thrusting (e.g., Wynne-Edwards 1972; Sangster and Bourne 1982; *see* Easton, this volume). To the north, the subprovince is bound by the Neoarchean plutonic and metasedimentary rocks of the Opatica Subprovince (Card and Ciesielski 1986). The western Abitibi Subprovince in Ontario contains the following supracrustal-dominated domains: 1) the Abitibi greenstone belt (which is divided into a northern and southern region (Dimroth et al. 1982, 1984); 2) Batchawana greenstone belt; 3) Benny, Hutton and Parkin greenstone belts; and 4) the Temagami greenstone belt. Greenstone belts listed in items 2 through 4, above, occur to the south of the Abitibi greenstone belt, and their relationship to the Abitibi greenstone belt is not well known. Granitoid plutonic and/or gneiss-dominated domains within the Abitibi Subprovince consist of discrete batholiths within the greenstone belts (e.g., in Ontario, these are the Kenogamissi, Watabeag and Round Lake batholiths) and large areas of plutonic and gneissic rock between greenstone belts (e.g., the Ramsey–Algoma granitoid complex; *see* Figure 11.2).

The Michipicoten greenstone belt, commonly considered at least a temporal equivalent of the Abitibi greenstone belt (Percival and Card 1983; Jackson and Sutcliffe 1990), is described by Williams, Stott, Heather et al. (this volume). The Ramsey–Algoma granitoid complex is described with the Abitibi Subprovince. Its relationship to the other elements of the Abitibi Subprovince is not well known and it may, like the Pontiac Subprovince, warrant subprovince status.

In this chapter, the geology of the western Abitibi Subprovince is reviewed through a systematic description of its supracrustal assemblages, intrusions and structures; most of the text is concerned with the description of supracrustal assemblages. Each assemblage is described separately to avoid making, or inferring, correlations that may not be valid. Note that, in general, contact relationships are only mentioned where they are known or some inference can be made as to their nature. These descriptions serve as the basis for discussions, found at the end of this chapter, of greenstone belt assemblages, structure, magmatism, mineralization, and tectonic models for the southern Abitibi greenstone belt and adjacent components of the central Superior Province.

Mineralization in the Western Abitibi Subprovince

In this chapter, the mineral deposits are referred to in the context of the supracrustal assemblage that encloses them. The genesis of the deposits is not discussed; however, where available, references that discuss the deposits are provided. A summary of the main deposit types and supracrustal assemblages that contain them is given in Figure 11.3. The main mineral deposit types in the western Abitibi Subprovince include: volcanic-associated, massive, base metal sulphide (VMS), lode gold, komatiite-associated Ni-Cu-PGE and iron formation. Examples of assemblages that contain these deposits include:

VMS: Adair (Normetal Mine in Quebec), Kidd–Munro, Kamiskotia, Blake River and Geneva assemblages

Lode gold: Detour, Tisdale, Three Nations and Timiskaming assemblages

Komatiite-associated Ni-Cu-PGE: Kidd–Munro, Eldorado and Bartlett assemblages

Iron formation: Boston, Hanrahan, Marion, Eldorado and Chambers–Briggs assemblages

ABITIBI GREENSTONE BELT

The Abitibi greenstone belt (e.g., Goodwin and Ridler 1970) is unique amongst greenstone belts of the Canadian Shield in that it: has a high ratio of supracrustal to intrusive rocks; is the largest greenstone belt in the world; has a generally low metamorphic grade; and contains a diverse spectrum of richly mineralized deposits.

Overview

HISTORICAL OVERVIEW

Geological mapping between approximately 1900 and 1970 established that the Abitibi greenstone belt consists mainly of mafic to felsic metavolcanic units, metasedimentary units, and a variety of granitoid rocks (*see* Thomson 1943, 1946, 1948, 1949, 1950; Thomson and Griffis 1944; Thomson et al. 1950). An early suite of metavolcanic rocks, the "Keewatin", was shown to have been deformed prior to, and after, the deposition of a younger suite of metasedimentary and alkalic metavolcanic rocks, the "Timiskaming". The regional extent of the "breaks", now known as the Larder–Cadillac shear zone and the Porcupine–Destor deformation zone[1], was established, as was their spatial relationship to gold mineralization. Significant deposits of iron (e.g., Adams and Sherman mines; Dubuc 1966) and copper-zinc mineralization (e.g., Kamiskotia camp and Kidd Creek Mine) were also discovered during this early work. Nickel mineralization associated with ultramafic flows was discovered in the Frederick House Lake area (Alexo and Dundonald deposits) and south of Timmins in the Shaw Dome area (Redstone, Hart (Tontine), McWatters, Langmuir No.1 and Langmuir No.2 deposits; *see* Fyon et al., this volume).

Between ~1970 and the early 1980s, systematic mapping continued, for example, in the Blake River group (*see* Jensen 1972, 1975a, 1975b, 1978a, 1978b, 1982, 1985a, 1985b, 1989; Jensen and Langford 1985) and the region between the Kenogamissi and Watabeag batholiths (*see* Pyke 1970, 1973, 1976, 1978, 1982; Pyke and Middleton 1971; Pyke et al. 1973). Significantly, the emphasis of this work was lithology and regional correlation of rock units. Chemical analysis of rock units became a routine component of geological studies and established the extent of the kom-atiitic, tholeiitic, calc-alkalic, and alkalic suites (e.g.,

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[1] *Shear or fault zone is used to refer to a discrete, mappable zone of high strain. Deformation zone is used to refer to a broader region of high strain consisting of fabrics, folds, faults and shears.*

Goodwin 1979; Jensen and Langford 1985; Pyke et al. 1973). Reflecting the approach of this era, contacts or transitions between the recognized stratigraphic "groups" were either established as, or inferred to be, stratigraphic or tectonically modified stratigraphic contacts. This "stratigraphic" view of the Abitibi greenstone belt attributed the recurrence of similar rock suites in the "stratigraphic column" to cyclic volcanism. The cycles were thought to consist of, from base to top, komatiitic, tholeiitic, calc-alkalic, and alkalic suites. As many as 4 cycles of volcanism (some incomplete) were postulated; however, only 2 cycles were thought to be areally significant (e.g., Jensen and Langford 1985).

From the mid–1980s to the present, systematic mapping has continued and specialized studies have been conducted (e.g., structural analysis of fault systems, geochemical analysis of the alteration and vein systems associated with gold mineralization, and strontium, neodymium and lead isotopic studies). Precise U-Pb geochronology has been fundamental to advancing the understanding of the evolution of the belt (e.g., Nunes and Jensen 1980; Nunes and Pyke 1980, 1981; Corfu et al. 1989, in press). The geochronological data indicate that similar rock suites were erupted at different times; but, it has also revealed "out-of-sequence" stratigraphy suggesting tectonic juxtaposition of units. Recent structural studies have demonstrated the presence of thrust faults in the greenstone belt. In addition, radiogenic isotope studies have revealed that some magmatic elements of the belt are mantle derivatives, whereas others may have an older evolved crustal component. As a result of these advances, the classical stratigraphic view of the Abitibi greenstone belt is changing and more work is required to more fully understand the significance of, and relationship between, the distinct supracrustal assemblages that comprise the greenstone belt.

Currently, there is much interest in development of evolutionary models for the Abitibi greenstone belt. Some notable overviews include those of Jensen (1985a), Dimroth et al. (1983b) and Hodgson and Hamilton (1989). These conceptual models help to identify the main problems to be resolved concerning the evolution of the greenstone belt. A discussion of these models is given at the end of this chapter.

GEOLOGICAL OVERVIEW

The Abitibi greenstone belt (*see* Figures 11.2 and 11.4) is bounded to the: 1) south by the Ramsey–Algoma granitoid complex, the Cobalt Embayment and the Pontiac Subprovince; 2) north by the Archean Opatica Subprovince (*see* Williams, this volume); 3) west by the Ivanhoe Lake cataclastic zone, which marks the western boundary of the Kapuskasing Structural Zone (e.g., Percival and Card 1983, 1985); and 4) east by the Proterozoic Grenville Province.

Historically, the western limit of the Abitibi greenstone belt has been associated with a north-trending chain of Archean granitoid complexes including the Kenogamissi Batholith (*see* Figure 11.2) and complexes west and northwest of Timmins (e.g., MERQ–OGS 1983; *see* Figure 11.2). The boundary between these granitoid complexes and the southern Abitibi greenstone belt is spatially associated with the northern extension of faults related to the western margin of the Cobalt Embayment (Jackson and Sutcliffe 1990, Figure 1). These faults are generally east-side-down normal structures with a component of apparent sinistral horizontal displacement. Consequently, the region west of these faults, that extends northward from the western margin of the Cobalt Embayment, may simply represent a deeper erosion level of a once-continuous greenstone belt extending at least to the Ivanhoe Lake cataclastic zone.

The Abitibi greenstone belt consists of a northern ("internal") belt and a southern ("external") belt (*see* Figure 11.4; Dimroth et al. 1983b). The northern belt is characterized by abundant tonalite-trondhjemite-granodiorite intrusions, large anorthosite complexes, paucity of ultramafic flows, and greenschist or higher regional metamorphism (Dimroth et al. 1983b). The southern belt is characterized by fewer tonalite-trondhjemite-granodiorite intrusions, abundant ultramafic flows, and greenschist or lower regional metamorphism (Dimroth et al. 1983b). In Ontario, only a limited part of the northern Abitibi greenstone belt is preserved in the Detour and Burntbush areas, north of the Lake Abitibi Batholith (*see* Figures 11.4 and 11.5). The following overview is mainly concerned with the southern Abitibi greenstone belt. Recent summaries of the Abitibi greenstone belt in Ontario include those of Jensen and Langford (1985) and Hodgson (1983a).

In the southern Abitibi greenstone belt, metavolcanic and associated metasedimentary units, and synvolcanic peridotitic to granodioritic intrusions formed between ~2.75 and ~2.70 Ga (Corfu et al. 1989). Relatively old rocks, such as the ~2.8 billion-year-old (Mortensen 1987b) felsic metavolcanic rocks of the Obatagamau Formation near Chibougamau (*see* Figure 11.4), have not been found in the southern Abitibi greenstone belt. Between ~2.70 and ~2.68 Ga, large volumes of foliated tonalite-granodiorite batholiths were emplaced followed by more massive granodiorite, granite, feldspar ± quartz porphyry, and syenite bodies (Corfu et al. 1989). During and subsequent to this magmatism, alluvial-fluvial clastic metasedimentary rocks (Hyde 1980) and alkalic metavolcanic rocks (Cooke and Moorhouse 1969) formed and are now in spatial association with regional, steeply dipping shear zones (Porcupine–Destor deformation zone and Larder–Cadillac shear zone). Regional deformation occurred prior to, and after, formation of the alluvial-fluvial clastic metasedimentary rocks and associated alkalic metavolcanic rocks. Metamorphic grade within the supracrustal rocks is generally subgreenschist to greenschist facies and rises to amphibolite facies near some intrusions (Jolly 1978, 1980).

Bedding and tectonic fabric in the southern Abitibi greenstone belt generally dip steeply to moderately (90° to 45°); however, shallow dips are present in some areas such as the core of the Blake River group (Jensen 1975a, 1975b), southwest of Timmins (Pyke 1982; Piroshco and Kettles 1987), in the northern Swayze area (Milne 1972) and the area adjacent to the Kenogamissi Batholith (Milne and Breaks 1972; Bright 1984). Folds are generally east or west trending and upright. There are south-verging thrust faults that predate and postdate deposition of the younger

		ADAIR	SCAPA	STEELE	STOUGHTON-ROQUEMAURE	DUFF-COULSON-RAND	KIDD-MUNRO	KAMISKOTIA	CARSCALLEN	HOYLE	TISDALE	KRIST	PORCUPINE	THREE NATIONS	GARRISON	DELORO	ELDORADO	BARTLETT	PETERLONG	BOWMAN	GEIKIE	WATABEAG	HALLIDAY	KINOJEVIS NORTH	KINOJEVIS SOUTH	BLAKE RIVER	GAUTHIER
	Age (Ma) (1)	2730					2717	2707				2700								2705	2700			2700	2700	2700	2700
UNITS	Fe Formation (2)		X		O/S	X	S	S	O	X	X	S	X	X		X	O,S	O,S						X	X	X	X
	Ultramafic Flows	X	X	X	P	X	P	X	X	X	P	X	X	X		X	P	P	mi	P				X	X	X	X
	Ultramafic–Mafic Intrusions	X	X	X	P	mi	P	P	P	X	X	X	X	X		mi	P	P					mi	mi	mi	mi	X
	Mafic Flows	P	X	P	P	X	P	P	P	X	P	X	X	X					P	P	P		X	P	P		X
	Intermediate–Felsic Fragmental Rocks	P	X	X	X	P	mi	P		X	X	P	X	X		mi	P	P		X		P	P	X	X	P	P
	Intermediate–Felsic Effusive Flows	P	X	X	X		X	P		X	X	X	X	X		P			X			P		X	X	P	
	High–silica Rhyolite		X				P	P		X	X	X	X	X													
	Alkalic Volcanic Rocks		X	X	X	X	X	X	X	X	X	X	X	X		X	X	X	X	X		X	X	X	X	X	X
	Conglomerate (3)		M	X	X	M	X	X	X	M	X	X	M	M/Cs		X	X	X	X	X		X	X	X	X	X	X
	Sandstone (4)		T	X	X	T	X	X	X	T	X	X	T	T/Cb		X	X	X	X	X		X	X	X	X	X	X
MINERAL DEPOSITS	Volcanogenic Massive Sulphides	P	X	X	X	X	P	P	X	X	X	X	X	X	X	X				mi				X	X	P	
	Komatiite–associated Ni–Cu–PGE	X	X	X	X	X	P	X	X	X	X	X	X	X		X	P	P				X	X	X	X	X	X
	Gold	X	X	X	X	X	X	mi	X	P	P	X	P	P	P	X	X						X	mi	mi	X	mi
STRUCTURAL STYLE	Homoclinal	Y	Y		X		X	Y			X	X	X	Y					Y	Y				Y			U
	Folded				Y		P	X		Y	Y	Y	Y	X		Y	Y				Y		Y		Y	Y	U
CONTACT RELATIONSHIPS	Base (5)	I	F?													U	D?/F	I	I	F	F?	U	U	F	U	F/D?	U
	Top (5)	F?	F													F/D	D?/F	F	I/U	D?/F	U	U	U/F?	F/D?	F/D?	U	U
	North (5)			F	F	F	F/U		U/F	U/F	U			D/F													
	South (5)			F	F/U	F	U/F		U/F	U/F	F			F													
AEROMAGNETIC PATTERN	Flat Aeromagnetic Pattern (6)					L	I				I	L							I			L	L			L	L
	Variable Aeromagnetic Pattern	Y		Y	Y		Y				Y					Y	Y	Y		Y	Y			Y	Y		
	Assemblages Description (7)	AB	C1	A5	A1	B	A2	A2	A5	C1	A3/5	B1	C1	C2	C2	B2	A6	A6	A5	A3	A4	B	B	A4	A4	B	B1

Figure 11.3. Summary of the main features of supracrustal assemblages of the southern Abitibi greenstone belt.

		TIMISKAMING	HEARST	WHITNEY	MIDLOTHIAN	CATHARINE-PACAUD	SKEAD	MCELROY	LARDER LAKE	BOSTON	RIDOUT	MUSKEGO-REEVES	HANRAHAN	HORWOOD	RANEY-NEWTON	HALCROW-SWAYZE	MARION	GARNET-TOOMS	HONG KONG	GRIFFIN	DISMAL	WART	GENEVA	BLUEWATER	CHAMBERS-BRIGGS	ARSENIC	COMMAND
	Age (Ma) (1)	2685 2675	2700 2680				2700		2705						2700					2729	2711 2698				2737		
UNITS	Fe Formation (2)	X	O	O		X	X	X		O	O		O				Cf/O		O		O,S	O	X	X	O/S	X	X
	Ultramafic Flows	X	X	X	X	mi	X	mi	P	P	X	P	X	X	X	P	X	P		X	X	X	X	X	X	mi	X
	Ultramafic–Mafic Intrusions	X	X	X	X	mi	mi	P	mi		X		P							X	X	mi			P	P	
	Mafic Flows	X	X	X	X	P	X	P	P	P	X	P	X	P		P	X	P	P	P	P	X	mi	P	P	P	P
	Intermediate–Felsic Fragmental Rocks	X	X	X	X	X	P	mi	X	P	X	P	P	X	P		P	mi		mi	P	X	P	X	P	P	X
	Intermediate–Felsic Effusive Flows	mi?	X	X	X	mi		X	X		X	P	mi		P		P				P	X	P	X	P	P	X
	High-silica Rhyolite	X	X	X	X						X											X			X	P	X
	Alkalic Volcanic Rocks	P	X	X	X	X	X	X	X			X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X
	Conglomerate (3)	Cs	M		P	X		X			M/Cs	M	X	X	M	X	X	X	X	X	X	P		X	M	M	X
	Sandstone (4)	Cb	T	P	Cb	X		X	M		T/Cb	T	X	X	T	X	X	X	X	X	X	P		X	T	T	X
MINERAL DEPOSITS	Volcanogenic Massive Sulphides	X	X	X	X						X	X	X	P?	X	X	X	X	X	X	X	X	P	X	X	X	X
	Komatiite-associated Ni–Cu–PGE	X	X	X	X		X	X			X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X
	Gold	P	P		mi	mi	mi	X	mi				X				X	X	X	X	X	X	X	X	X	mi	X
STRUCTURAL STYLE	Homoclinal	Y		X	Y	Y	Y	Y				X	Y	X	Y	X	Y	X	X	X	Y	X			X	Y	
	Folded		Y	Y					Y	Y		Y	X		X	Y	X	Y		Y		Y			Y	X	
CONTACT RELATIONSHIPS	Base (5)	D/F	F	U	U/D?	I	D/F	F?													I						
	Top (5)	U	U	U	U	D/F	F?	F													F						
	North (5)								F	F	U/F	I		U	U	F/D		U/F	D/U	F		I/F	I/U	I	I	F/I	F/U
	South (5)								F	F	U/F	U		U	F/D	F/U		D/U	I	F/I		F	I	I/U	F	F	U
AEROMAGNETIC PATTERN	Flat Aeromagnetic Pattern (6)						L					I		I	L	I		I	I								
	Variable Aeromagnetic Pattern	Y	Y	Y	Y	Y		Y	Y	Y		Y				Y		Y	Y								
	Assemblages Description (7)	C2	C1	C1	C2	A5	B1	A5	A3	A6	C2	ABC	A6	A5	B	A6	A6	A6	A5	A5	A6	C1	B1	A5	A6	A2	A5

Notes:

a) "mi" indicates minor presence; blank field indicates indeterminate
b) "X" denotes feature that has low, or no, probability of being present
c) "P" indicates that feature is present; "Y" indicates that the assemblage displays a particular structural style or aeromagnetic pattern
d) "?" indicates uncertainty in observation

1. *Dates are approximate. Where two dates are given, the age of the assemblage is bracketed between the two dates.*
2. *O Oxide facies; S Sulphide facies; Cf Carbonate facies*
3. *M Matrix supported; Cs Clast supported*
4. *T Turbidites; Cb Cross-bedded units (fluvial?)*
5. *Where base and top of assemblages cannot be determined with confidence, the north and south contacts are described.*
   D Depositional; F Fault; I Intrusive; U Unknown
6. *I Intermediate intensity; L Low intensity*
7. *Assemblage description codes correlate with those in Table 11.1. Assemblages that are similar to one another have the same code.*

Figure 11.3. Summary of the main features of supracrustal assemblages of the southern Abitibi greenstone belt.

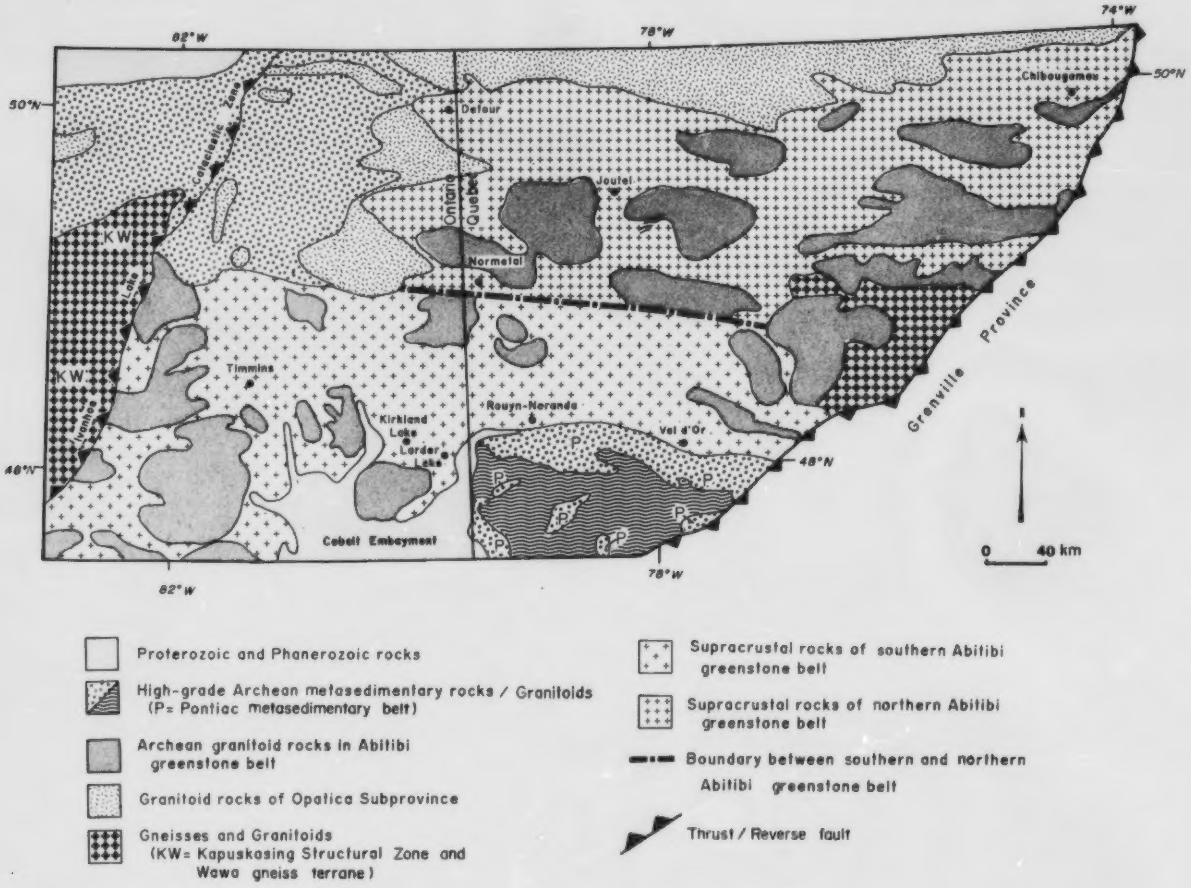
74°W
78°W
82°W
50°N
48°N
Chibougamau
Defour
Joutel
Normetal
Ontario
Quebec
Cataclastic Zone
KW
Lake
Ivanhoe
Timmins
Kirkland Lake
Larder Lake
Rouyn-Noranda
Val d'Or
Cobalt Embayment
P
Grenville Province
0 40 km
Proterozoic and Phanerozoic rocks
High-grade Archean metasedimentary rocks / Granitoids (P= Pontiac metasedimentary belt)
Archean granitoid rocks in Abitibi greenstone belt
Granitoid rocks of Opatica Subprovince
Gneisses and Granitoids (KW= Kapuskasing Structural Zone and Wawa gneiss terrane)
Supracrustal rocks of southern Abitibi greenstone belt
Supracrustal rocks of northern Abitibi greenstone belt
Boundary between southern and northern Abitibi greenstone belt
Thrust / Reverse fault

Figure 11.4. The Abitibi greenstone belt. Simplified from MERQ–OGS (1983) and Bedrock Geology of Ontario Map 2543 (*see* map case). Division between northern and southern Abitibi greenstone belt *after* Dimroth et al. (1982).

alluvial-fluvial metasedimentary and alkalic metavolcanic rocks (Thomson 1943, 1948; Hewitt 1963; Piroshco and Kettles 1987, 1988).

Major gold camps are spatially associated with steeply dipping shear zones, such as the Larder–Cadillac shear zone and the Porcupine–Destor deformation zone which transect the belt for over 300 km in a general easterly direction (Dimroth et al. 1983a; Hodgson 1986; Colvine et al. 1988). In general, steeply dipping foliations and steeply plunging stretching lineations and minor folds characterize these zones, and both dextral and sinistral horizontal components of displacement are reported (e.g., Larder–Cadillac shear zone; Toogood and Hodgson 1985, 1986; Hamilton 1986; Tourigny et al. 1989; Robert 1989, 1990). The shear zones are spatially associated with the youngest supracrustal and plutonic assemblages and, in part, deform these assemblages. These observations indicate that the shear zones were active late in the history of the greenstone belt; an earlier history is difficult to substantiate, but has been suggested (e.g., Dimroth et al. 1982, 1983b). Correlation of rock assemblages across these shear zones has not been unequivocally demonstrated. The Larder–Cadillac shear zone, in part, marks the boundary between the southern Abitibi greenstone belt and the Pontiac Subprovince (*see* MERQ–OGS 1983).

Seismic reflection profiles from the southern Abitibi greenstone belt (*see* Figure 11.5) indicate that regional subhorizontal features are an important component of the Abitibi greenstone belt (Green et al. 1990; Jackson, Sutcliffe et al. 1990). Many of the reflectors are laterally continuous beneath complex surface geology, wherein units are moderately to steeply dipping. One of the subhorizontal reflectors (a-a′ of Figure 11.5b) appears to truncate other reflectors (b-b′ of Figure 11.5b) and may represent a subhorizontal fault zone (Jackson, Sutcliffe et al. 1990). Beneath the surface trace of some fault zones, reflectors are truncated to depths of 15 km (*see* Figure 11.5; e.g., the Porcupine–Destor deformation zone and the Mulven fault and Misema–Mist fault, the latter of which may be an extension of, or a fault related to, the Kirkland Lake fault).

Detailed Assemblage Descriptions

The most recent regional descriptions of supracrustal rocks in the Ontario part of the Abitibi greenstone belt are those of Pyke (1982), Jensen (1985a) and Jensen and Langford (1985). From these descriptions and resulting analysis, a regional stratigraphic framework for the greenstone belt was proposed. Geochronology has shown that some of the age relationships predicted by the stratigraphic model are incorrect. Consequently, the interpretation of contact relationships between supracrustal units is approached cautiously; in particular, attention is drawn to data that are inconsistent with unmodified stratigraphic contacts. This approach should identify regions where further study might clarify the nature of the supracrustal contact relationships, thereby improving our understanding of the belt. The supracrustal rocks of the Abitibi Subprovince are described in terms of assemblages, a term defined in the chapter entitled Archean of Ontario: Introduction.

In many cases, the assemblages described correspond with regional units identified by previous workers. In these instances, the original name of the unit is retained, but in general, the status of the unit is changed, e.g., Timiskaming group (old name) has become Timiskaming assemblage (new name). The main features, locations and contact relationships of the assemblages discussed in this chapter are given in Figures 11.6, 11.7 and 11.8 (*see also* Figure 11.3). Assemblages are, for convenience, described according to the following geographic areas (*see* Figure 11.6): Detour Lake area, Burntbush area, Cochrane–Lake Abitibi area, Belford–Strachan area, Swayze area, Watabeag area, Round Lake area and Shining Tree area.

The geophysical character of major units forms part of the descriptions. Extensive use was made of regional scale airborne electromagnetic (AEM) and magnetic surveys. The abbreviations for assemblages on Figure 11.7 correspond to the abbreviations used on the tectonic map (*see* Map 2577, map case).

DETOUR LAKE AREA

Rocks of the Detour Lake area (Johns 1982; Marmont 1987) are bounded to the north by gneissic and plutonic rocks of the Opatica Subprovince and to the west by the Hopper Lake granitoid complex (massive to weakly foliated quartz diorite, quartz monzonite and granodiorite; Marmont 1987). The greenstone belt is here subdivided into the Detour, Lower Detour and Vandette assemblages. The temporal and stratigraphic relationships between these assemblages is not known. The metamorphic grade of the belt is generally lower amphibolite facies, although towards the Quebec border there is an apparent decrease to greenschist facies in the Detour assemblage (Marmont 1987). A thin, upper amphibolite facies contact metamorphic aureole is developed adjacent to granitoid intrusions external to the greenstone belt (Marmont 1987). Feldspar and/or amphibole-porphyritic, trondhjemite, granodiorite, monzonite and aplite-pegmatite intrusions occur as dikes and as small stocks, which cut the supracrustal rocks (Marmont 1987). Felsic dikes near the Detour Mine area are ~2722 Ma (Marmont and Corfu 1989).

Detour Assemblage

The Detour assemblage (DT; *see* Figure 11.7) consists of iron-rich tholeiitic basalt and associated gabbro, diorite and quartz diorite stocks and dikes, and minor east-striking, discontinuous, chemical metasedimentary units (identified by AEM conductors) along the southern edge of the assemblage. Both north- and south-facing units exist (Johns 1982; Marmont 1987). Basaltic rocks consist of massive, pillowed, and feldspar-megacrystic flows, autoclastic flow breccia and tuff breccia. Ultramafic rocks, of uncertain origin, occur only in the southern part of the assemblage, near the Detour deformation zone (Marmont 1987).

A regional penetrative foliation is northeast- to east-striking. The Detour deformation zone, an east-southeast-striking, ductile shear zone has been traced for about 20 km, from the Hopper Lake granitoid complex, to the central part of the belt (Marmont 1987). Moderately west-plunging stretching lineations and kinematic asymmetry are consistent with north-side-up dextral displacement (Marmont 1987). At its eastern limit of exposure, the Detour deformation zone approaches the unexposed contact between the Detour and Lower Detour assemblages.

Lower Detour Assemblage

The poorly exposed Lower Detour assemblage (LD) is an easterly trending domain (*see* Figure 11.7). Argillite and polymictic conglomerate appear to predominate in the south, and quartzofeldspathic and chloritic wackes appear to predominate in the northern part of the assemblage (Marmont 1987). Metasedimentary rocks in the southern part of the assemblage face north (Marmont 1987).

Vandette Assemblage

Features of the Vandette assemblage (VE; *see* Figure 11.7) at the southern part of the belt are interpreted largely from drill core and aeromagnetic and airborne electromagnetic (AEM) data (OGS 1989). Steeply dipping, south-facing, iron-rich tholeiitic basalt units predominate in the northern part of the assemblage. These units include: massive, pillowed and feldspar-megacrystic flows, and autoclastic flow breccia, tuff and tuff breccia (Johns 1982). Felsic tuffs occur in 2 bands near the central and southern parts of the assemblage (Johns 1982; Marmont 1987). Conductive chemical metasedimentary rocks occur locally within the assemblage (Johns 1982).

The main structures within this assemblage include a fold with a north-striking axial trace interpreted from the arcuate trace of AEM conductors (OGS 1989) and an east- to southeasterly striking ductile shear zone (Cassiopea–Westmin deformation zone) that occurs close to the contact between the Vandette and Lower Detour assemblages (Marmont 1987). This shear zone is interpreted to have a dextral component of slip and occurs where back-to-back facing directions between the Vandette and Lower Detour assemblages are reported (Marmont 1987).

Economic gold concentrations are only reported from the Detour Mine (Marmont 1986) within the Detour assemblage. Mineralization at the mine is associated with quartz veins and zones of sulphide impregnation, localized in sheared mafic and ultramafic metavolcanic rocks within the Detour deformation zone (Marmont 1986, 1987).

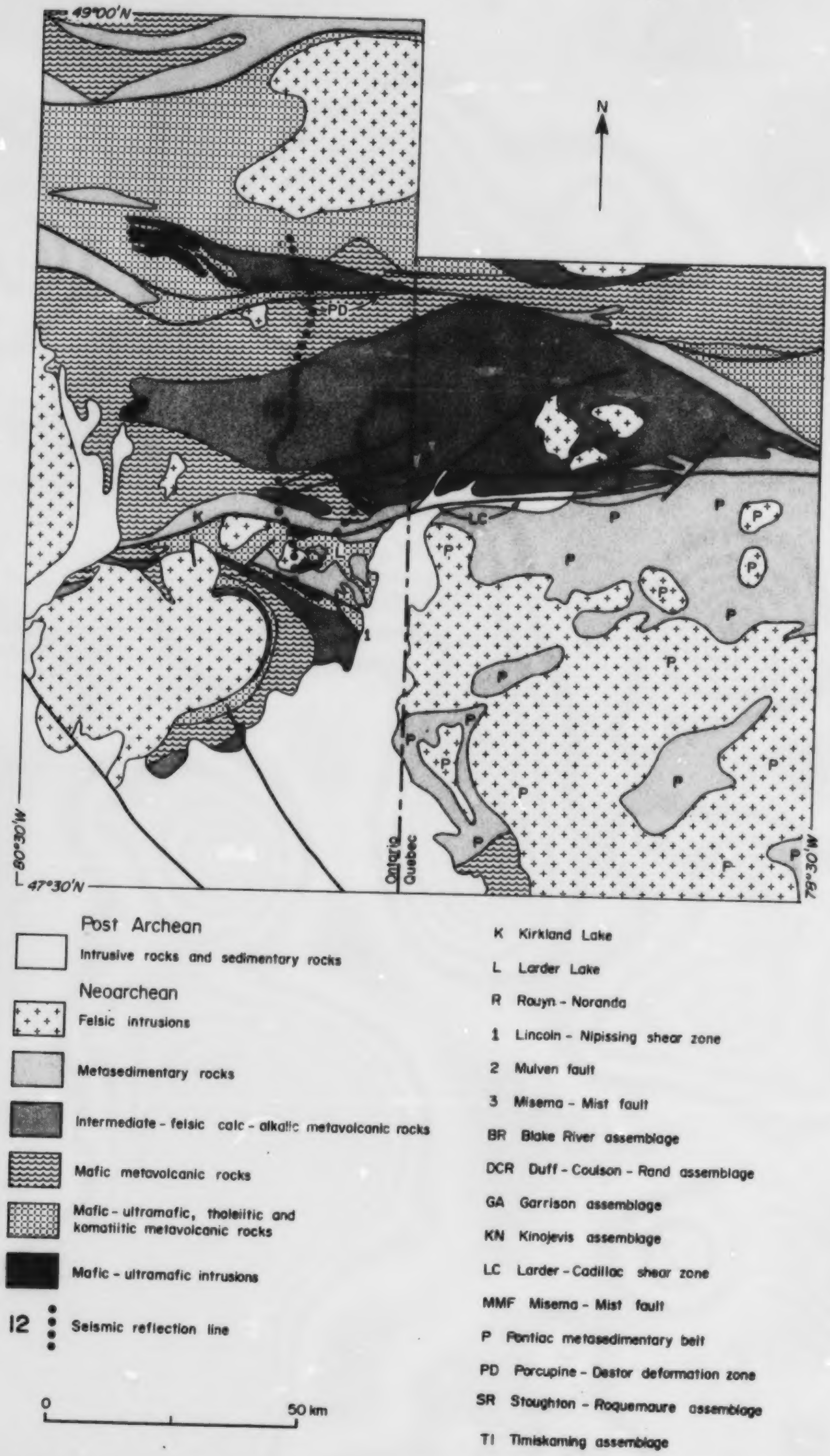
49°00'N
N
PD
K
LC
L
P
1
80°30'W
47°30'N
Ontario
Quebec
79°30'W
Post Archean
Intrusive rocks and sedimentary rocks
Neoarchean
Felsic intrusions
Metasedimentary rocks
Intermediate - felsic calc - alkalic metavolcanic rocks
Mafic metavolcanic rocks
Mafic - ultramafic, tholeiitic and komatiitic metavolcanic rocks
Mafic - ultramafic intrusions
12 Seismic reflection line
0
50 km
K Kirkland Lake
L Larder Lake
R Rouyn - Noranda
1 Lincoln - Nipissing shear zone
2 Mulven fault
3 Misema - Mist fault
BR Blake River assemblage
DCR Duff - Coulson - Rand assemblage
GA Garrison assemblage
KN Kinojevis assemblage
LC Larder - Cadillac shear zone
MMF Misema - Mist fault
P Pontiac metasedimentary belt
PD Porcupine - Destor deformation zone
SR Stoughton - Roquemaure assemblage
TI Timiskaming assemblage

Figure 11.5. Seismic reflection profiles that reveal regional subhorizontal elements in the subsurface of the southern Abitibi greenstone belt, *modified from* Green et al. (1990) and Jackson et al. (1990). **a)** Location of seismic reflection lines 12 and 12A that are shown in b and c, respectively.

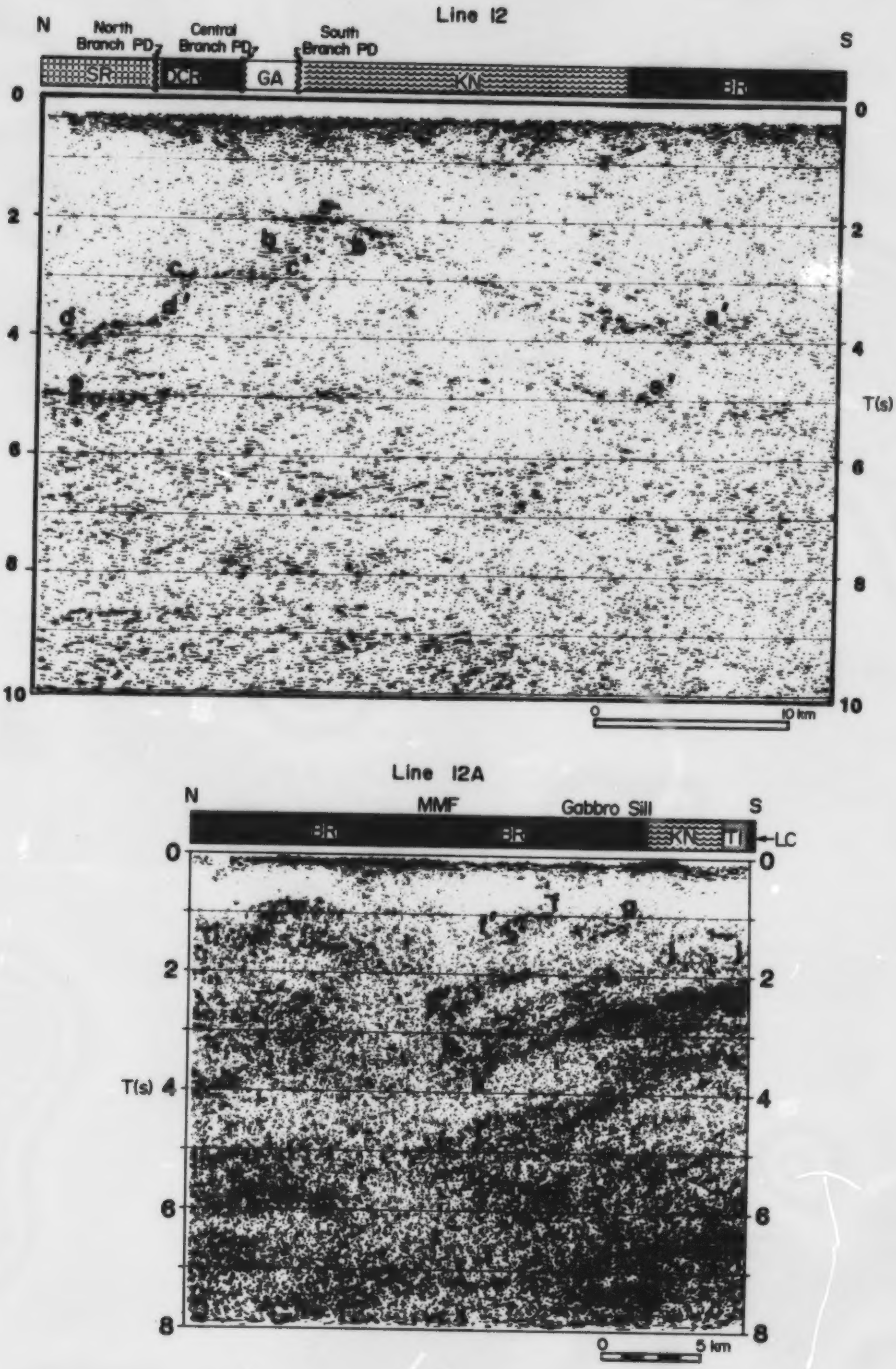

Figure 11.5. Seismic reflection profiles that reveal regional subhorizontal elements in the subsurface of the southern Abitibi greenstone belt, *modified from* Green et al. (1990) and Jackson et al. (1990). b) Northern half of seismic reflection line 12. Note that reflector a-a′ truncates reflector b-b′, suggesting that a-a′ may be a subhorizontal fault and/or shear zone. Note also that the Porcupine–Destor deformation zone and its splays appear to penetrate to at least 15 km below the surface of the Earth. c) Seismic reflection line 12A showing "stacked" subhorizontal reflectors. Reflection f-f′ may correlate with either a gabbro sill or a shear zone located south of the sill. *See* Jackson, Sutcliffe et al. (1990) for further details. T(s) = two-way travel time in seconds. For b and c, horizontal:vertical scale is ~1:1, and 1s = ~3 km.

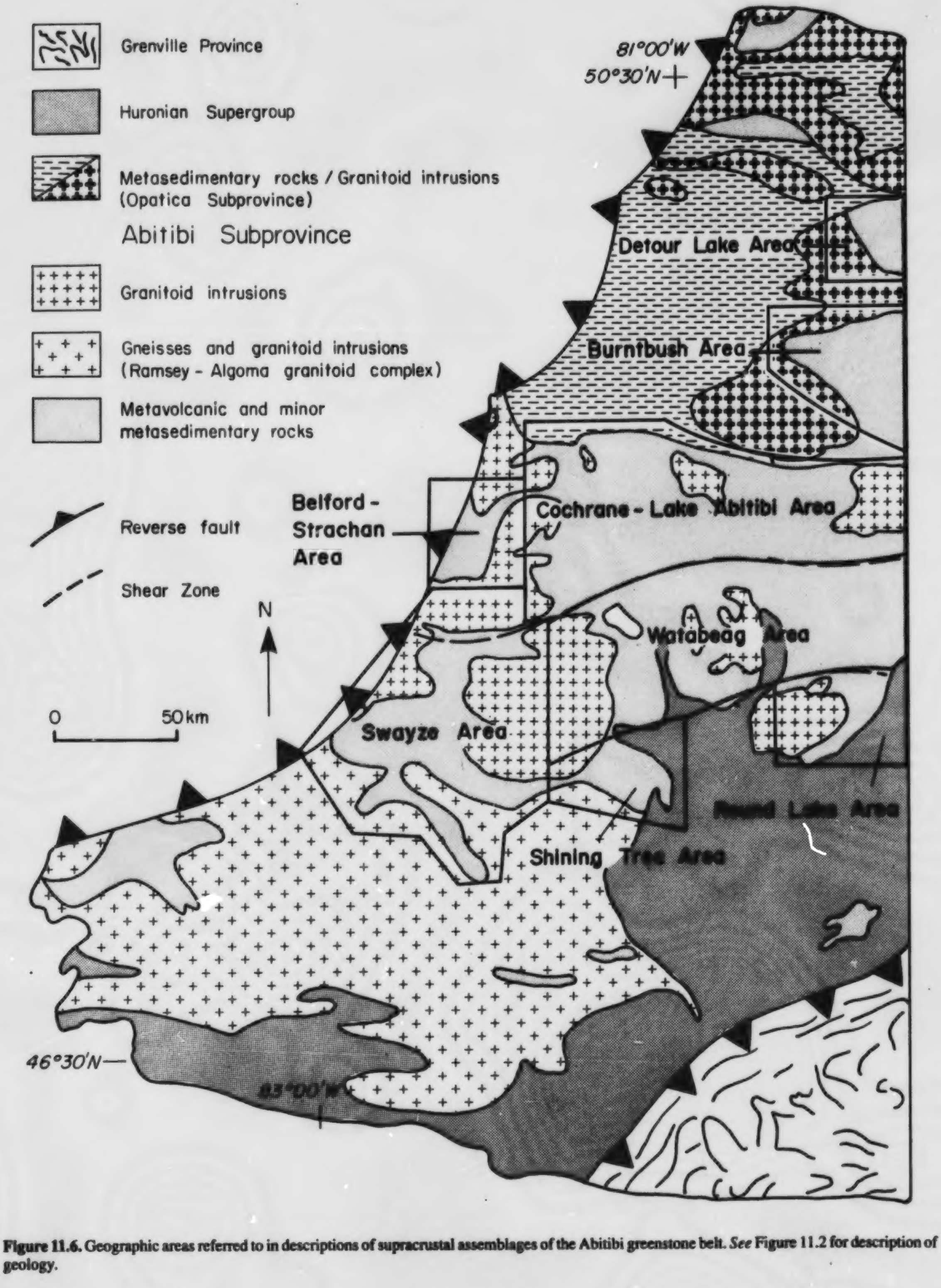
Grenville Province
Huronian Supergroup
Metasedimentary rocks / Granitoid intrusions (Opatica Subprovince)
Abitibi Subprovince
Granitoid intrusions
Gneisses and granitoid intrusions (Ramsey - Algoma granitoid complex)
Metavolcanic and minor metasedimentary rocks
Reverse fault
Shear Zone
N
0
50 km
81°00'W
50°30'N
Detour Lake Area
Burntbush Area
Belford - Strachan Area
Cochrane - Lake Abitibi Area
Watabeag Area
Swayze Area
Round Lake Area
Shining Tree Area
46°30'N
83°00'W

Figure 11.6. Geographic areas referred to in descriptions of supracrustal assemblages of the Abitibi greenstone belt. *See* Figure 11.2 for description of geology.

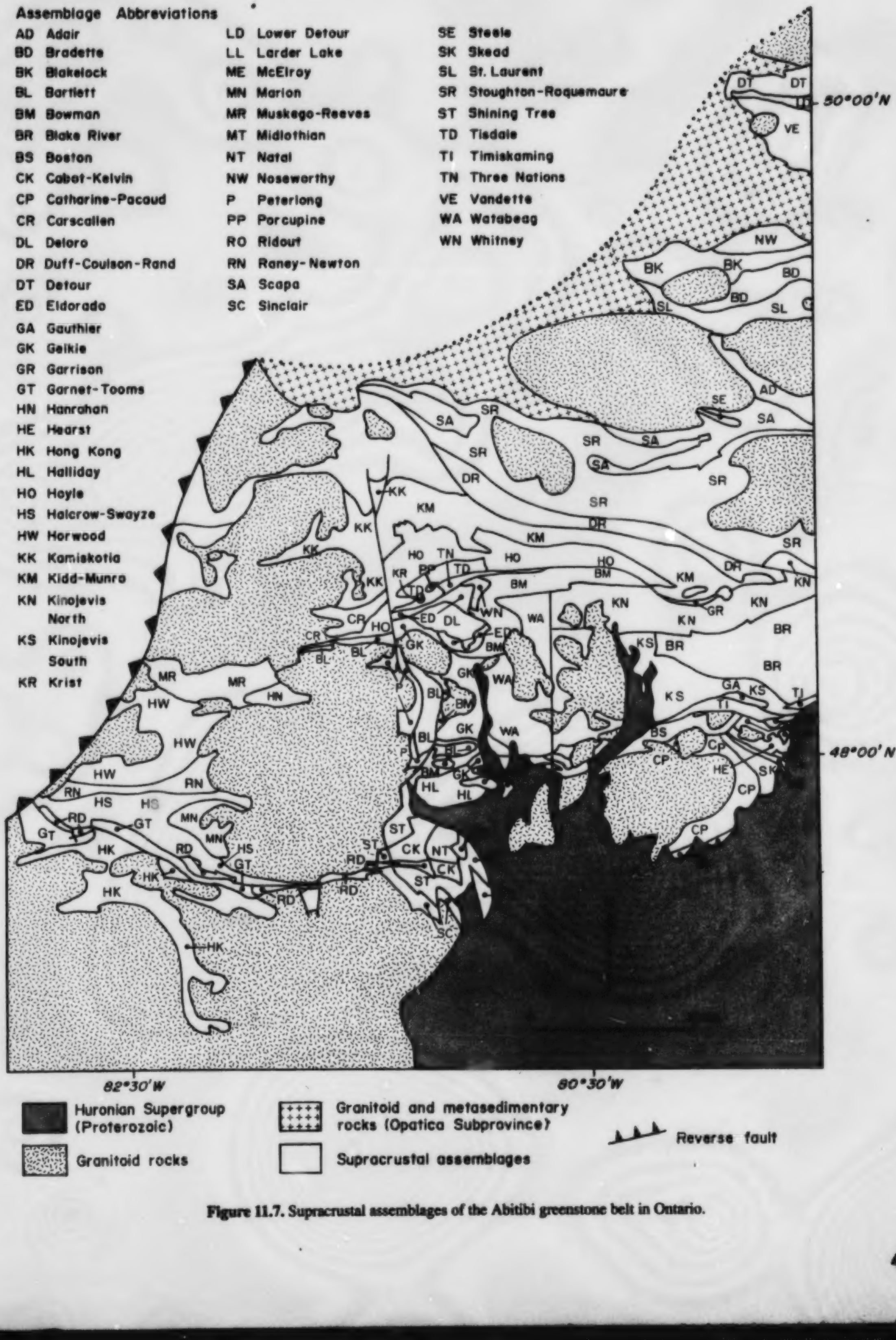
Assemblage Abbreviations
AD Adair
BD Bradette
BK Blakelock
BL Bartlett
BM Bowman
BR Blake River
BS Boston
CK Cabot-Kelvin
CP Catharine-Pacaud
CR Carscallen
DL Deloro
DR Duff-Coulson-Rand
DT Detour
ED Eldorado
GA Gauthier
GK Geikie
GR Garrison
GT Garnet-Tooms
HN Hanrahan
HE Hearst
HK Hong Kong
HL Halliday
HO Hoyle
HS Halcrow-Swayze
HW Horwood
KK Kamiskotia
KM Kidd-Munro
KN Kinojevis North
KS Kinojevis South
KR Krist
LD Lower Detour
LL Larder Lake
ME McElroy
MN Marion
MR Muskego-Reeves
MT Midlothian
NT Natal
NW Nosewerthy
P Peterlong
PP Porcupine
RO Ridout
RN Raney-Newton
SA Scapa
SC Sinclair
SE Steele
SK Skead
SL St. Laurent
SR Stoughton-Roquemaure
ST Shining Tree
TD Tisdale
TI Timiskaming
TN Three Nations
VE Vandette
WA Watabeag
WN Whitney
50°00'N
48°00'N
82°30'W
80°30'W
DT
VE
NW
BK
BD
SL
AD
SE
SA
SR
DR
KK
KM
HO
TN
PP
TD
KR
BM
KN
GR
BR
CR
DL
ED
WN
WA
GK
KS
GA
TI
BS
CP
HE
SK
BL
MR
HN
HW
RN
HS
RD
GT
MN
HK
ST
CK
NT
HL
P
SC
Huronian Supergroup (Proterozoic)
Granitoid rocks
Granitoid and metasedimentary rocks (Opatica Subprovince)
Supracrustal assemblages
Reverse fault

Figure 11.7. Supracrustal assemblages of the Abitibi greenstone belt in Ontario.

BURNTBUSH AREA

Assemblages of the Burntbush area (*see* Figure 11.6; Lumbers 1962, 1963; Wilson 1979; Johns 1982) are bounded to the north and west by a massive biotite-granodiorite intrusion and to the southwest by the "Case batholith" (Bennett et al. 1966). To the south, the assemblages are bounded between the Mistawak batholith on the north and metasedimentary rocks of the Scapa assemblage. To the east, the assemblages continue into Quebec. Although uncommon, Late Precambrian quartz diabase dikes of the north-striking Matachewan dike swarm and northeast-striking Abitibi dike swarm are present (Fahrig and West 1986).

The following assemblages are recognized: Noseworthy, Blakelock, Bradette, St.Laurent and Adair. The distribution of units and structural discontinuities within the assemblages are interpreted on the basis of airborne geophysical data (OGS 1989) and drill core data.

Noseworthy Assemblage

The east-trending, cuspate Noseworthy assemblage (NW) extends from the Quebec–Ontario border to the northwestern limit of the greenstone belt (*see* Figure 11.7). The assemblage extends to the east into Quebec, where it corresponds, in part, to the Casa Berardi deformation zone (Avramtchev and LeBel-Drolet 1981, Map 305). The assemblage consists largely of metasedimentary rocks dominated by turbidites (Wilson 1979; Johns 1982).

	Duff–Coulson–Rand	Garrison	Hoyle	Kamiskotia	Kidd–Munro	Krist	Porcupine	Stoughton–Roquemaure	Three Nations	Tisdale
Carscallen			F	F						
Duff–Coulson–Rand					F			F		
Garrison					F?					
Hoyle				F	S?/F				$\overset{\leftarrow}{U}$	S?/F?
Kamiskotia					F					
Kidd–Munro										
Krist							S			S
Porcupine									U↑	
Stoughton–Roquemaure										
Three Nations										$\overset{\leftarrow}{U}$

Key to Figure 11.8

F – Fault contact
S – Stratigraphic contact
U – Unconformity or disconformity (arrows point to younger assemblage)
I – Assemblages separated by an intrusion
X – Contact not determined
Blanks indicate assemblages not in contact

Assemblage pairs in fault contact along Porcupine–Destor deformation zone include:

Hoyle/Bowman
Kidd–Munro/Kinojevis North
Kidd–Munro/Garrison*
Garrison*/Kinojevis North
Three Nations/Bowman
Three Nations/Whitney
Tisdale/Eldorado
Carscallen/Bartlett
Hoyle/Bartlett
* Enclosed by Porcupine–Destor deformation zone

Figure 11.8. Summary of contact relationships between supracrustal assemblages of the southern Abitibi greenstone belt.

	Blake River	Boston	Bowman	Catharine–Pacaud	Deloro	Eldorado	Gauthier	Geikie	Halliday	Hearst	Kinojevis South	Kinojevis North	Larder Lake	Midlothian	McElroy	Peterlong	Skead	Timiskaming	Watabeag	Whitney
Bartlett			F/X					F						X		I/X				
Blake River											F	F								
Boston				F						F			F		F		F	F		
Bowman					X	F/←S?		X	F?/X							F			F/X	X
Catharine–Pacaud																	U↕/F			
Deloro						S↕/X														
Eldorado																				
Gauthier												X						U/F		
Geikie									X					U?↕/X						
Halliday														U?↕						
Hearst													U?/X				F?	X		
Kinojevis South												F?						X	F	
Kinojevis North																		U↕	I	
Larder Lake															F		F	F		
Midlothian																				
McElroy																	X			
Peterlong																				
Skead																				
Timiskaming																			X	
Watabeag																				

	Halcrow–Swayze	Hanrahan	Hong Kong	Horwood	Marion	Muskego–Reeves	Raney–Newton	Ridout
Garnet–Tooms			S/F					F
Halcrow–Swayze					S?		S?/F	F
Hanrahan						S?		
Hong Kong								
Horwood						X	S?/F	
Marion								
Muskego–Reeves								
Raney–Newton								

Figure 11.8. Summary of contact relationships between supracrustal assemblages of the southern Abitibi greenstone belt.

Blakelock Assemblage

The east-striking Blakelock assemblage (BK; *see* Figure 11.7) has a uniformly high aeromagnetic signature and numerous AEM conductors, and is therefore interpreted to consist predominantly of mafic metavolcanic rocks, interlayered with magnetite-rich and/or sulphide-rich or carbonaceous chemical metasedimentary rocks. Pillow basalt in the western part of the assemblage faces north; elsewhere, few facing data exist (Wilson 1979; Johns 1982). A contact strain aureole around the northwestern margin of the Bateman pluton (Wilson 1979) may be indicated by the presence of moderately foliated and well-banded, folded rocks, interpreted by Wilson (1979) to be tuffaceous and pyroclastic basaltic rocks.[2] A poorly exposed band of conglomerate-bearing wacke occurs in the western part of the assemblage (Wilson 1979) and has a low magnetic signature (OGS 1989).

Felsic metavolcanic rock may underlie the eastern part of the assemblage (Johns 1982). Alternatively, like the western part of this assemblage, the high magnetic signature and the presence of many AEM conductors may indicate that the eastern part of the assemblage consists mostly of basaltic rock interlayered with magnetite-rich iron formation.

Bradette Assemblage

The Bradette assemblage (BD; *see* Figure 11.7) consists of calc-alkalic dacitic and rhyolitic, quartz-phyric tuff, lapilli tuff, pyroclastic breccia, tuff breccia and flows interlayered with carbonaceous metasedimentary units (Johns 1982). On the basis of its low aeromagnetic signature and numerous AEM conductors, this assemblage has been extrapolated westward into Tomlinson Township, south of the Bateman pluton, which is a massive to weakly foliated, biotite granodiorite to granite (Wilson 1979).

At the eastern edge of the Bradette assemblage, a southeast-striking, sinistral shear zone, near the interface between the Blakelock and Bradette assemblages, is inferred on the basis of deflected east- and southeast-striking AEM conductors (OGS 1989). This inferred shear zone may represent the continuation of the Casa Berardi deformation zone recognized in Quebec, which is locally associated with significant gold mineralization (Pattison et al. 1986).

St. Laurent Assemblage

The St. Laurent assemblage (SL; *see* Figure 11.7) consists of iron- and magnesium-rich, tholeiitic basalts and andesites which are massive, pillowed, feldspar-megacrystic, or fragmental. Minor ultramafic units are also present (Wilson 1979; Johns 1982). The assemblage is characterized by a uniformly high magnetic signature, and numerous west- to northwest-striking, continuous, AEM conductors (OGS 1989), which are interpreted to correspond to metasedimentary rocks. Feldspar-megacrystic flows, of probable tholeiitic affinity, occur in the Twopeak Lake area, south and west of the Bateman pluton (Wilson 1979). Pillowed basalts in the western part of the assemblage dip steeply and face south. Pillowed basalts in the eastern part of the assemblage, north of the Mistawak batholith face north and east (Johns 1982).

The contact between the St. Laurent and Blakelock assemblages is interpreted to be a structural discontinuity on the basis of: 1) an angular discordance in the strike of AEM conductors between the assemblages; and 2) "moderately foliated and well banded", folded, tuffaceous basalts, which are localized at the junction of the assemblages (Wilson 1979; *see* footnote [2]).

Adair Assemblage

The Adair assemblage (AD) trends southeasterly along the southern boundary of the Burntbush area (*see* Figure 11.7) and consists of south-facing, steeply dipping, tholeiitic and calc-alkalic metavolcanic rocks (Lumbers 1963; Johns 1982). Iron- and magnesium-rich tholeiitic basalts predominate in the southern half of the assemblage. Basaltic metavolcanic rocks include massive, amygdaloidal, pillowed and coarse-grained (1 to 3 cm) feldspar-megacrystic flows, interlayered with minor, mafic, interflow breccia (Lumbers 1963).

The northern half of the assemblage consists predominantly of intermediate and felsic, quartz- and feldspar-phyric, calc-alkalic massive flows and fragmental rocks (Lumbers 1963; Johns 1982). Coarse-grained felsic breccia is rare, but is concentrated at the southeastern end of the assemblage, near the Quebec border (Johns 1982; Lumbers 1963). Similar felsic metavolcanic rocks, along strike near Normetal, Quebec (e.g., Avramtchev and LeBel-Drolet 1981, Map 308), are ~2726 Ma (Mortensen 1987a).

The northwestern part of the assemblage may consist of interlayered wacke, calc-silicate units, felsic metavolcanic rocks and iron formation (Johns 1982). These units of grunerite-chert, magnetite-rich, sulphide-rich, and carbonaceous chemical metasedimentary rocks (Lumbers 1963) are marked by northwest-striking AEM conductors (OGS 1989) and could be part of the Scapa assemblage.

Penetrative foliation within the assemblage strikes northwesterly, parallel to stratigraphic units (Johns 1982). A dextral(?) shear zone, identified at the Quebec–Ontario boundary, strikes west to northwest and affects both the Adair assemblage and the northern edge of the Patten River stock (Lumbers 1963; Avramtchev and LeBel-Drolet 1981, Maps 305 and 308). The associated shear foliation may be related to the northwest-striking foliation present throughout the assemblage (Lumbers 1963; Johns 1982).

Regional metamorphic conditions of upper greenschist-lower amphibolite facies prevail. The granodiorite-granite Case and Mistawak batholiths have amphibolite facies contact aureoles (Lumbers 1962, 1963; Johns 1982).

No significant mineralization is known from the Ontario part of the Adair assemblage; however, the felsic metavolcanic rocks of the assemblage extend into Quebec where they host volcanogenic base metal massive sulphides at the Normetal Mine (Avramtchev and LeBel-Drolet 1981, Map 308).

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[2] *In many greenstone belts, rocks previously mapped as "tuffs" are now either interpreted as highly strained supracrustal rocks, commonly including pillowed basalts, or the status as "tuffs" is equivocal. Interestingly, "tuffs" are commonly reported from high-strain environments such as contact strain aureoles adjacent to plutons and from shear zones that transect supracrustal assemblages.*

ASSEMBLAGES NORTH OF THE ABITIBI BATHOLITH

The 5 to 12 km wide, 100 km long clastic metasedimentary Scapa assemblage coincides with the boundary between the "internal" (north) and "external" (south) zones of the Abitibi greenstone belt (*see* Figures 11.4 and 11.7; Dimroth et al. 1982). This linear metasedimentary assemblage is at least spatially associated with a structural discontinuity that separates the internal and external zones of the Abitibi greenstone belt (Dimroth et al. 1982). Linear domains of metasedimentary rocks and regional structural discontinuities represent a common lithostructural arrangement along subprovince boundaries and within subprovinces of the Superior Province.

Scapa Assemblage

The Scapa assemblage (SS; *see* Figure 11.7) consists of south-facing, steeply dipping, turbidite metasedimentary rocks interlayered with amphibole-rich, calc-silicate rocks and a thin unit of mafic metavolcanic rock (Lumbers 1962, 1963; *see* Steele Assemblage). The Scapa assemblage extends westward from the Quebec–Ontario border to the western limit of the Abitibi greenstone belt. In part, the assemblage defines the contact between the gneissic and plutonic rocks of the Opatica Subprovince and the northern part of the Abitibi greenstone belt (Cochrane–Lake Abitibi area; *see* Figure 11.6). The assemblage is 12 km wide north of the Lake Abitibi Batholith, but ~60 km to the east in Quebec, the assemblage is less than 5 km wide (Lumbers 1963; Avramtchev and LeBel-Drolet 1981, Map 308). Although a shear zone occurs within the Adair assemblage near the contact with the Scapa assemblage, the contact between the Adair and Scapa assemblages is not exposed.

Cross-bedding, channelling, local unconformities and intraformational breccias occur in turbidite beds close to the southern boundary, along Lake Abitibi (Lumbers 1962, p.22). Narrow (less than 10 cm) calc-silicate layers, consisting of grunerite-quartz, with traces of hornblende, pyrite and pyrrhotite (Lumbers 1962), occur within turbidite and conglomerate beds (Lumbers 1962, 1963). Rare conglomerate layers, exposed along the north shore of Lake Abitibi, contain well-rounded quartz- and feldspar-porphyritic boulders (Lumbers 1963). Lamprophyre dikes cut the metasedimentary rocks (Lumbers 1962).

An east- to east-southeast-striking, steeply dipping penetrative foliation is well developed throughout the assemblage (Lumbers 1962, 1963). Regionally continuous, dextral(?) shear zones occur within and along the southern boundary of the assemblage. These shear zones are marked by well-developed foliation, z-shaped minor folds and carbonate alteration. At the southern limit of the assemblage, a high-strain zone separates the Scapa assemblage and metavolcanic rocks to the south (Bonis "sequence" of the Stoughton–Roquemaure assemblage; *see* Stoughton–Roquemaure Assemblage). This high-strain zone is marked by contorted and folded bedding, faulted quartz veins and zones of pyritic, chlorite-carbonate alteration.

Regional metamorphism attained upper greenschist to lower amphibolite facies, with garnet-bearing and amphibole-hornfels zones present adjacent to the Scapa pluton (Lumbers 1962, 1963).

Steele Assemblage

The southeasterly trending Steele assemblage (SE; *see* Figure 11.7) is less than 2 km in width and consists of mafic metavolcanic flows and minor interlayered metasedimentary rocks, cut by quartz- and feldspar-porphyritic felsic dikes (Lumbers 1962; Johns 1982). The assemblage is structurally interleaved with the Scapa assemblage (Lumbers 1962) and extends for 10 km east-southeasterly from the Case batholith to a northeast-striking diabase dike, where it terminates.

The mafic metavolcanic rocks include amygdaloidal, pillowed, massive and feldspar-porphyritic basaltic flows (some of tholeiitic affinity) and mafic pyroclastic deposits. The feldspar-phyric flows that occur along the southern edge of the assemblage may, based on mineralogical descriptions, be calc-alkalic in composition. These flows are in contact with metasedimentary rocks of the Steele assemblage (Johns 1982). Thin units of biotite-quartz-microcline-sphene-garnet schists of possible sedimentary protolith are interlayered with the basalts (Lumbers 1962; Johns 1982).

Numerous west-northwest-striking, sulphide-bearing shear zones cut the assemblage (Lumbers 1962). The largest shear zones conform with the southern and northern contacts of the Scapa assemblage (Lumbers 1962, p.39, 41). The deflection of bedding and foliation within the Scapa assemblage and the presence of z-shaped minor folds within the shear zones is consistent with an oblique, dextral component of slip. Northwest- and northeast-striking faults, possibly conjugates, also occur in this area. The northeast-striking set consistently shows sinistral offset, whereas the northwest-striking set shows dextral offset (Lumbers 1962). The relationships between these faults and the larger-scale, east- to northeast-striking ductile shear zones is unclear.

The metamorphic grade of the Steele assemblage is amphibolite facies, except adjacent to the Case batholith, where amphibole-hornfels conditions were attained (Lumbers 1962).

A columbite-tantalite-molybdenite-spodumene-bearing pegmatitic dike in the Case batholith, near the contact with the Scapa and Steele assemblages (Lumbers 1962, p.28), represents the only known mineralization within the area.

ASSEMBLAGES IN THE COCHRANE–LAKE ABITIBI AREA

The Cochrane–Lake Abitibi area (*see* Figure 11.6) is informally subdivided into the Stoughton–Roquemaure, Duff–Coulson–Rand, Kidd–Munro, Kamiskotia, Carscallen, Tisdale, Krist, Porcupine, Three Nations and Garrison assemblages.

Stoughton–Roquemaure Assemblage

The Stoughton–Roquemaure assemblage (SR; *see* Figure 11.7) includes most of the "Stoughton–Roquemaure group" of Jensen and Langford (1985). The assemblage is bounded to the: 1) north by the Scapa assemblage; 2) south by the Bradburn–Coulson shear zone and the Porcupine–Destor deformation zone; and 3) east by the Lake Abitibi Batholith (*see* Figures 11.7 and 11.9). The assemblage consists predominantly of komatiitic and magnesium- and iron-rich tholeiite (Eakins 1972; Jensen 1978a).

South of Lake Abitibi, the Stoughton–Roquemaure assemblage is south facing. It consists of peridotitic and basaltic komatiite, magnesium- and iron-rich tholeiite, finely layered chert, iron formation and felsic tuff horizons. All of these units are interlayered with coarse-grained, gabbroic,

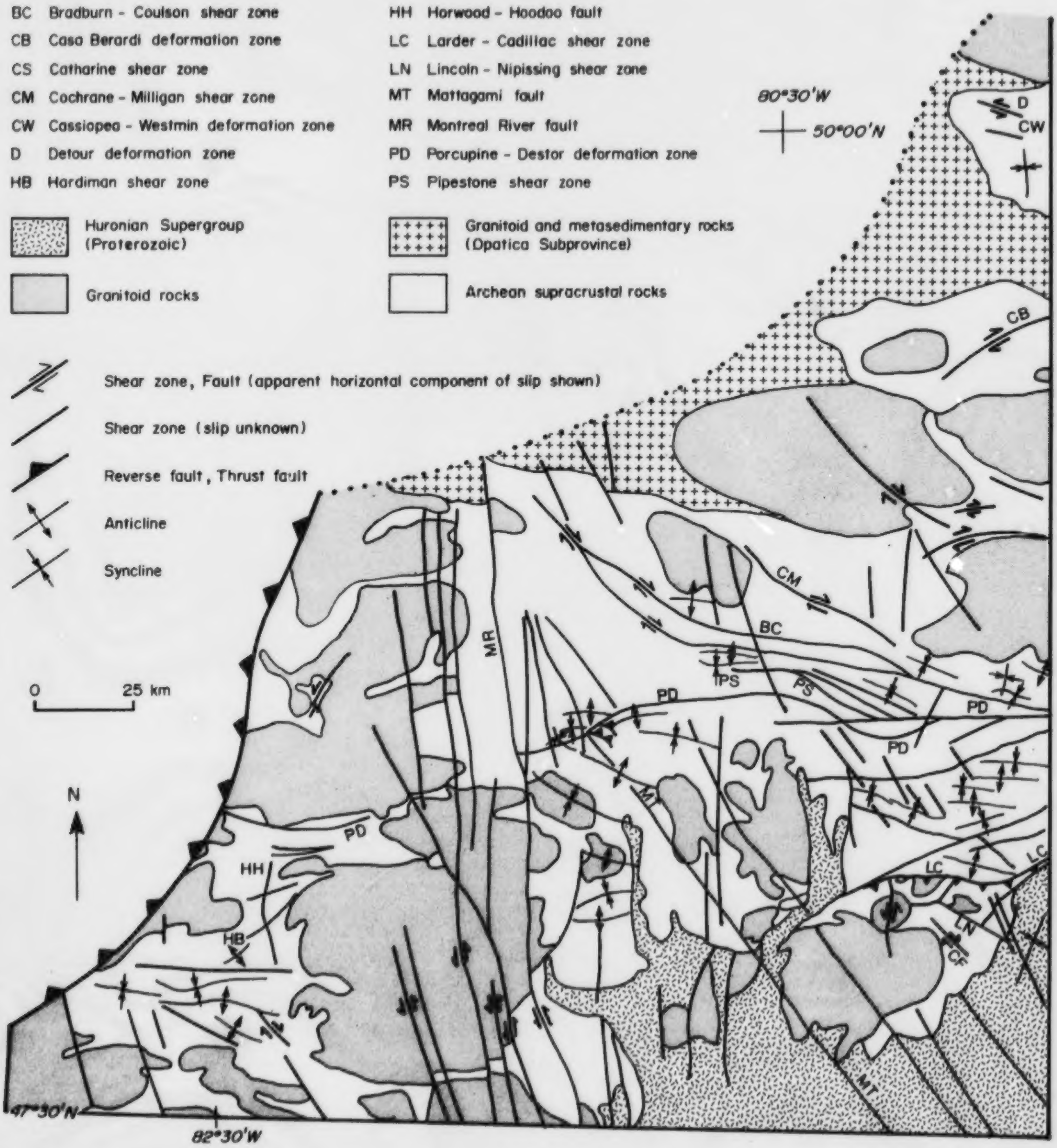
BC Bradburn - Coulson shear zone
CB Casa Berardi deformation zone
CS Catharine shear zone
CM Cochrane - Milligan shear zone
CW Cassiopea - Westmin deformation zone
D Detour deformation zone
HB Hardiman shear zone
HH Horwood - Hoodoo fault
LC Larder - Cadillac shear zone
LN Lincoln - Nipissing shear zone
MT Mattagami fault
MR Montreal River fault
PD Porcupine - Destor deformation zone
PS Pipestone shear zone
Huronian Supergroup (Proterozoic)
Granitoid and metasedimentary rocks (Opatica Subprovince)
Granitoid rocks
Archean supracrustal rocks
Shear zone, Fault (apparent horizontal component of slip shown)
Shear zone (slip unknown)
Reverse fault, Thrust fault
Anticline
Syncline
0
25 km
N
80°30'W
50°00'N
47°30'N
82°30'W
D
CW
CB
CM
BC
PS
PD
MR
HH
HB
LC
LN
CF
MT

Figure 11.9. Simplified structural geology of the Abitibi greenstone belt in Ontario.

peridotitic and dunitic intrusions or thick tabular flows (Jensen and Langford 1985). Several units consist of basal peridotitic komatiite, overlain successively by basaltic komatiite, magnesium-rich tholeiitic basalt, and iron-rich tholeiitic basalt (Jensen and Langford 1985). Within these thicker units, ultramafic and basaltic-komatiitic flows generally decrease in thickness upward, whereas magnesium-rich tholeiitic basalt flows increase in thickness stratigraphically upwards (Jensen and Langford 1985). These repetitive sequences vary from 300 to 1700 m in thickness, show various states of completeness (Jensen and Langford 1985) and are generally either pillowed or massive (Eakins 1972; Jensen and Langford 1985). A 350 m thick iron-rich tholeiitic basalt occurs at the interpreted base of the assemblage (Jensen and Langford 1985).

Immediately north of the Lake Abitibi Batholith, the Stoughton–Roquemaure assemblage consists of north-facing, steeply dipping mafic metavolcanic rocks called the Bonis "sequence" by Lumbers (1962, 1963). This "sequence" consists of porphyritic, amygdaloidal, massive and pillowed units (Lumbers 1962) of probable tholeiitic affinity, interlayered with numerous, continuous, conductive interflow metasedimentary units (OGS 1989).

Adjacent to the northern contact of the Lake Abitibi Batholith, upper greenschist and lower amphibolite facies metavolcanic rocks bear a well-developed east- to northeast-striking foliation or gneissosity (Lumbers 1962). This part of the sequence is structurally juxtaposed against the south-facing, metasedimentary rocks of the Scapa assemblage (Lumbers 1962). Based on indirect evidence, such as back-to-back younging, sheared and folded lithologies and carbonatization, the contact between the Bonis "sequence" and the Scapa assemblage is interpreted as a shear zone (Lumbers 1962). Minor z-shaped folds along this shear zone are consistent with a dextral component of horizontal slip.

A dextral shear zone, herein called the Cochrane–Milligan shear zone, strikes eastward along the contact between the metavolcanic rocks of the Cochrane–Lake Abitibi area and the migmatized metasedimentary rocks of the Opatica Subprovince. Just east of a large granitoid intrusion at the western edge of the Stoughton–Roquemaure assemblage, the Cochrane–Milligan shear zone strikes southeast towards the Porcupine–Destor deformation zone, with which it may merge south of the Lake Abitibi Batholith. The trace of the Cochrane–Milligan shear zone is interpreted from deflected AEM conductors (OGS 1984). South of the Cochrane–Milligan shear zone, pillow basalts generally face to the south, whereas to the north, pillow basalts face north.

Duff–Coulson–Rand Assemblage

The Duff–Coulson–Rand assemblage (DR; *see* Figure 11.7) is fault bounded, narrow (less than 5 km) and extends northwesterly from the Porcupine–Destor deformation zone towards Cochrane. The extent of the assemblage is defined by its very low magnetic signature in comparison to the signature of adjacent assemblages. The northwestern boundary is defined by the merging of the bounding faults (*see* Figures 11.7 and 11.9).

Within the Duff–Coulson–Rand assemblage, there are 3 distinct rock types. The Rand unit consists of steeply dipping, generally south-facing, calc-alkalic andesite, dacite and rhyolite tuff and tuff breccia, interlayered with amygdaloidal, massive and pillowed calc-alkalic basalt and andesite (Jensen and Langford 1985). The Duff–Coulson–Rand assemblage occurs between the north and middle branches of the Porcupine–Destor deformation zone and was previously correlated with the type 2730 million-year-old Hunter Mine group in Roquemaure Township, Quebec (Dimroth et al. 1973; Jensen and Langford 1985). A 2715 Ma age (Corfu et al. 1989) for felsic metavolcanic rocks of the Rand unit, however, precludes such a correlation. In Rand Township, a local concentration of andesite, dacite and rhyolite breccia, intruded by numerous dikes and sills of porphyritic andesite, dacite and rhyolite, has been interpreted as an explosive volcanic vent (Jensen and Langford 1985). A strong penetrative, steeply dipping foliation is present in the unit (Jensen and Langford 1985).

The poorly exposed Coulson unit consists of well-bedded, steeply dipping, east-striking, clastic meta-sedimentary rocks in southern Coulson Township (Leahy and Ginn 1961a, 1961b). L.S. Jensen (Ontario Geological Survey, personal communication, 1989) suggests that this unit may contain a large component of tuffaceous material, equivalent to the calc-alkalic metavolcanic rocks exposed along strike, to the east, in Rand Township.

The Duff unit, in the northwestern part of the assemblage, consists of steeply dipping, northwest-striking and south-facing felsic metavolcanic rocks (based on exploration drilling; A.H. Green, Falconbridge Limited, personal communication, 1990). Other rock types include dacitic fragmental deposits, hornblende-porphyritic andesite flows, basaltic flows and interflow chemical metasedimentary units.

The contacts of the Duff–Coulson–Rand assemblage are not exposed; however, the deflection of AEM conductor traces suggests that the northwestern contacts are dextral shear zones. Similarly, the contact between the calc-alkalic metavolcanic rocks in Rand Township and metavolcanic rocks belonging to the Kidd–Munro assemblage to the south (*see* Kidd–Munro Assemblage) is also sheared (Satterly 1954).

Kidd–Munro Assemblage

The Kidd–Munro assemblage (KM) is an east- to southeast-striking assemblage (*see* Figure 11.7), which is bounded to the: 1) north by the Duff–Coulson–Rand assemblage; 2) west by the Mattagami fault (*see* Figure 11.9); and 3) south by the metasedimentary Hoyle assemblage (*see* Hoyle Assemblage). To the east, the assemblage terminates against the Porcupine–Destor deformation zone and the Stoughton–Roquemaure assemblage.

The assemblage consists of ultramafic, pyroxenitic, and basaltic komatiite, tholeiitic picrite, magnesium-rich tholeiite, high-alumina basalt, iron-rich tholeiite, icelandite (high-P_2O_5 andesite) and thin units of high-silica (FIII-type)

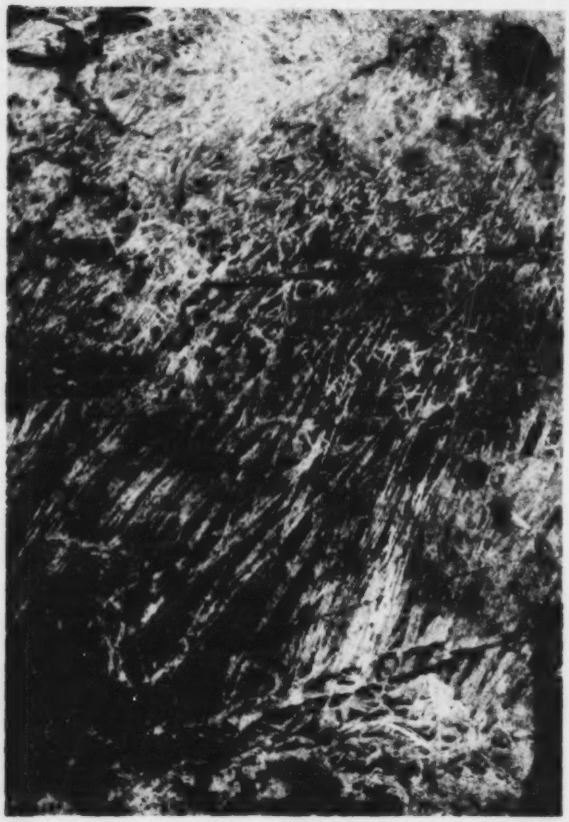

Figure 11.10. Olivine-spinifex textured part of a komatiite flow, Munro Township.

rhyolite.[3] Variolitic, massive and pillowed basalt units also occur (e.g., Leahy and Ginn 1961a, 1961b; Arndt and Nesbitt 1982). Associated with the metavolcanic units are layered, tholeiitic and ultramafic intrusions (e.g., Munro sill, and Centre Hill and McCool Hill complexes; MacRae 1969). Units generally have an east to southeast strike and steep dip. In the eastern part of the assemblage, east-striking folds and several east- to southeast-striking shear zones are present.

The classic ultramafic komatiites at Pyke Hill (Pyke et al. 1973) in Munro Township (Figure 11.10) are comprehensively described by Arndt (1976) and Arndt and Nesbitt (1982). Geochemical data suggest that 4 main types of basalts are present: Type 1—light rare earth element (LREE)-depleted, flat heavy rare earth element (HREE), low abundances of incompatible elements; Type 2—flat rare earth element (REE), enriched in incompatible elements; Type 3—Theo's flow basalts, strongly enriched in incompatible elements, slightly enriched LREE pattern, fractionated HREE; and Type 4—Warden Township basalts, having trace element characteristics similar in some respects to those of "Type 2" basalts (enriched incompatible elements, inferred flat REE patterns). The extent to which these local and detailed petrochemical relationships can be applied throughout the Kidd–Munro assemblage is not known.

Rhyolites and high-silica rhyolites, dated at 2715 Ma (Barrie 1990a; Barrie and Davis 1990) in the western part of the assemblage, near the Kidd Creek Mine, have relatively flat REE patterns ($[La/Yb]_n$ = 1 to 4), pronounced negative europium abundances (Eu/Eu^* = 0.20 to 0.61), low Zr/Y (2 to 6), high abundances of high field strength elements and low abundances of scandium and strontium (Lesher et al. 1986). Felsic metavolcanic rocks dating 2714±2 Ma from Beatty Township in the eastern part of the Kidd–Munro assemblage are chemically similar to rhyolites in the western part of Munro Township that are associated with komatiites. This suggests that komatiites within the Kidd–Munro assemblage are also about 2714 Ma.

The northern contact of this assemblage with the Duff–Coulson–Rand assemblage is not exposed, but locally appears to be a dextral shear zone (*see* Duff–Coulson–Rand Assemblage). The Pipestone shear zone (kinematics unknown) defines the contact with the Hoyle assemblage (*see* Hoyle Assemblage) to the south (e.g., Leahy and Ginn 1961a). It is not known if the irregularity of the western contact of the assemblage reflects a stratigraphic or structural interleaving of units.

At the western end of the Kidd–Munro assemblage, the giant, stratiform, polymetallic Kidd Creek deposit occurs within locally thickened (300 m), megascopic, S-folded, high-silica massive rhyolites to autobrecciated flows, and pyroclastic and effusive and intrusive mafic and ultramafic igneous rocks.

In the eastern part of the Kidd–Munro assemblage, in Munro Township, copper-zinc mineralization at the Potter Mine occurs within fragmental mafic rocks, associated with tholeiitic olivine basalts and picritic and peridotitic komatiites (Coad 1976). These metavolcanic rocks occur above and adjacent to the differentiated peridotite and/or gabbro of the tholeiitic Centre Hill Complex (Coad 1976). The contact between the Centre Hill Complex and the komatiitic flows to the west is a zone of shearing presumed to be a fault (e.g., MacRae 1969). Sulphide mineralization consists of massive and matrix pyrrhotite, sphalerite, chalcopyrite and minor pyrite (Coad 1976). Stringer sulphide mineralization is associated with a chlorite alteration zone (Coad 1976, p.144).

The following komatiite-associated, nickel-sulphide deposits occur in the Kidd–Munro assemblage, northeast of Timmins, in the Dundonald Township area: Alexo, Dundonald and Frederick House Lake. The area is underlain by mafic to felsic metavolcanic rocks, cut by the layered peridotite-gabbro Dundonald Sill (Naldrett 1964; Naldrett and Mason 1968; Muir and Comba 1979). Nickel sulphide mineralization, consisting of pyrrhotite, pentlandite, magnetite and chalcopyrite, occurs at the base of komatiite flows as: 1) massive stringers or nodules of pyrrhotite and pyrite and very fine-grained and disseminated pyrrhotite in graphite bands; 2) fine disseminated pyrrhotite (2 to 5% sulphides), most of which occurs in the basal cumulate part of the ultramafic flows; and 3) sulphide stringers or disseminated nodules which occur in serpentinized peridotite (Green and MacEachern 1990).

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[3] *See Basaltic Volcanism Study Project 1981, Chapter 1.2 for discussion of high-silica rhyolite.*

Asbestos has been mined from altered and deformed ultramafic rocks in Munro Township.

Kamiskotia Assemblage

The greenschist facies Kamiskotia assemblage (KK; *see* Figure 11.7) consists of the synvolcanic, 2707±2 million-year-old (Barrie and Davis 1990) tholeiitic Kamiskotia Gabbroic Complex, overlain by the 2705±2 million-year-old (Barrie and Davis 1990) Kamiskotia Volcanic Complex (Barrie 1990a). The Kamiskotia assemblage is bounded to the: 1) east by the north-northwest-striking, sinistral Mattagami fault; 2) north by an unnamed metavolcanic assemblage; and 3) west and south by the Carscallen assemblage (*see* Carscallen Assemblage). The western part of the Kamiskotia assemblage was strained and metamorphosed by the emplacement of 2696 to 2694 million-year-old hornblende-biotite tonalite to granite intrusions (Barrie 1990a). The igneous units in this assemblage dip near vertically and face both north and east (Barrie 1990a).

The Kamiskotia Gabbroic Complex is subdivided into 4 zones (Barrie 1990a): 1) a lower zone, consisting of partly layered, olivine-bearing cumulates along the southern and western margin; 2) a middle zone, consisting of gabbronorite and anorthositic gabbro-norite cumulates; 3) an upper zone, consisting of partly layered, ferroan gabbro-norite, anorthositic gabbro-norite and hornblende gabbro cumulates; and 4) granophyric rocks of intermediate and felsic composition above and along strike with the upper zone cumulates. The contact between the upper zone and granophyre is irregular. There are local stoped blocks of partly hybridized granophyric rock within the chloritized, quartz-rich upper zone gabbro (Hart 1984). Within the middle zone, a wide variety of mixed magma textures occur between gabbroic and tonalitic to granodioritic rocks.

The Kamiskotia Gabbroic Complex is overlain by, and is in part gradational with, genetically related rhyolites, basalts and minor evolved basalts and andesites of the Kamiskotia Volcanic Complex. The mafic metavolcanic rocks are divided into 4 divisions on the basis of P_2O_5, TiO_2, SiO_2 and Al_2O_3 content (Barrie 1990a): 1) tholeiites; 2) iron tholeiites; 3) icelandites[4]; and 4) high alumina basalts. Felsic volcanic rocks are rhyolite and high-silica rhyolite, having relatively flat REE patterns ($[La/Yb]_n$ = 1 to 4), pronounced depleted europium abundances (Eu/Eu^* = 0.20 to 0.61), elevated zirconium and yttrium abundances, low Zr/Y (2 to 6) and low abundances of scandium and strontium (Hart 1984; Lesher et al. 1986; Barrie 1990a).

Four hornblende-biotite tonalite to granite intrusions (2696 to 2694 Ma) were emplaced into the assemblage (Barrie and Davis 1990). These include a tonalite which exhibits mixed magma textures with Kamiskotia Gabbroic Complex rocks. This tonalite is believed to have intruded the base or the margin of the Kamiskotia Gabbroic Complex during crystallization, with limited magma mixing. A second tonalite emplaced into the assemblage has a well-developed foliation parallel to its margin. External to the assemblage are several discrete trondhjemite-tonalite to granodiorite-granite plutons. Amphibolite facies contact metamorphism accompanied the emplacement of these external plutons. North-northwest-striking Matachewan diabase dike swarms cut all rocks in the area.

Four volcanogenic copper-zinc ± gold ± silver (VMS) mines (Kam Kotia, Jameland, Canadian Jamieson and Genex) and several deposits occur within the Kamiskotia Volcanic Complex. The host rocks are predominantly mafic metavolcanic rocks, with subordinate high-silica rhyolite flows and tuffs (e.g., Lesher et al. 1986). Icelandite sills occur in proximity to the deposits (Binney and Barrie 1990). The deposits consist of several smaller lenses or masses of sulphide, generally within a restricted stratigraphic interval (less than 150 m) at each mine site (Binney and Barrie 1990); chloritic and sericitic wall rock alteration is typical (Binney and Barrie 1990). A close spatial, and possibly temporal, relationship is inferred to exist between shallow subaqueous to subaerial metavolcanic rocks and the ore deposits, suggesting that sulphide mineralization took place in a shallow, subaqueous regime (Binney and Barrie 1990). The sulphide minerals in the deposits are pyrite, pyrrhotite, chalcopyrite, sphalerite and minor magnetite; trace galena has been reported from the Kam Kotia mine (Pyke and Middleton 1971).

Although magmatic nickel-copper occurrences occur within the Kamiskotia Gabbroic Complex, no economic deposits are known.

Carscallen Assemblage

The upper greenschist to lower amphibolite facies Carscallen assemblage (CR; *see* Figure 11.7) is bounded to the: 1) west by 2696 to 2694 million-year-old hornblende-biotite tonalite to granite intrusions (Barrie and Davis 1990); 2) north by the Kamiskotia assemblage; 3) south by the Hoyle assemblage and the Porcupine–Destor deformation zone; and 4) east by the Mattagami fault. The assemblage consists predominantly of north-facing, steeply dipping, massive and pillowed basalt flows, showing both tholeiitic and calc-alkalic affinities, interlayered with minor amounts of high-silica rhyolite and oxide facies iron formation (Hawley 1927; Ferguson 1959; Barrie 1990a). Amygdaloidal basalt flows are rare (Ferguson 1959). An ultramafic intrusion occurs in the southern part of the assemblage. A 2687 million-year-old ultramafic lamprophyre suite, consisting of biotite lamprophyre, diopside-pyroxenite, and garnetite and garnet-rich feldspathic dikes (Ferguson 1959; Barrie 1990b), cuts the metavolcanic rocks.

Much of the assemblage lies within the Porcupine–Destor deformation zone, is highly strained and is characterized by an intense, penetrative flattening fabric and a 40° to 90° westerly plunging elongation lineation (Barrie 1990a). A 2687 million-year-old garnet-bearing ultramafic lamprophyre dike is deformed by the Porcupine–Destor deformation zone (Barrie 1990b).

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[4] *The term icelandite refers to a rock of intermediate composition having a P_2O_5 abundance exceeding 0.3 weight %. Two types of icelandites are recognized in the Abitibi greenstone belt: 1) TiO_2 less than 2.0 weight %; and 2) TiO_2 greater than 2.0 weight %.*

The contact between the Carscallen and Kamiskotia assemblages is marked by an increase in strain state toward the contact, suggesting that the assemblages are structurally juxtaposed. Rocks adjacent to the Carscallen–Hoyle assemblage contact are highly strained. Units on either side of the contact have a back-to-back facing relationship (Ferguson 1959; Barrie 1990a), which suggests a structural contact.

Hoyle Assemblage

The Hoyle assemblage (HO) trends easterly across the Cochrane–Lake Abitibi area (*see* Figures 11.6 and 11.7). It is bounded to the: 1) north by the Kidd–Munro assemblage; 2) south by the Tisdale and Three Nations assemblages and the Porcupine–Destor deformation zone; and 3) west by the Carscallen assemblage and the Mattagami fault. To the east, the assemblage terminates abruptly against the Kidd–Munro assemblage, possibly against a fault. It is dominated by wackes, interlayered with minor conglomerate and metavolcanic rocks (Hawley 1927; Harding and Berry 1939; Ferguson 1959). Carbonaceous metasedimentary rocks are particularly common near the contact with the metavolcanic rocks of the Tisdale assemblage located to the south (*see* Tisdale Assemblage).

Southwest of the Mattagami fault, the metasedimentary rocks have steep, southerly dips and are south facing (Hawley 1927; Harding and Berry 1939; Ferguson 1959). North- and south-facing turbidites, having strikes which range from east to north, occur in an isolated area in the central part of the assemblage. The abrupt change in facing direction over short distances (10 to 100s m) is consistent with the presence of tight folds in this area (e.g., Berry 1941; Ferguson 1964). In the eastern part of the assemblage (Carr and Beatty townships), steeply south-dipping, generally north-facing turbidite and arkosic beds occur (Prest 1952, 1957; Satterly 1952a). Bedded tuffs are also reported within this assemblage (Prest 1952, p.5).

Along the southern boundary, the Hoyle assemblage lies in contact with komatiitic and tholeiitic metavolcanic rocks of the Tisdale assemblage and clastic metasedimentary rocks of the Three Nations assemblage, and locally is the locus of the Porcupine–Destor deformation zone (Carr, Beatty and Thorneloe townships; Harding and Berry 1939; Prest 1952). Ultramafic fragmental rocks, of possible turbiditic origin, occur along and within the Porcupine–Destor deformation zone in the southwestern part of the Hoyle assemblage (e.g., Thorneloe Township; Fyon 1980; Choudry 1989). It is not certain if these ultramafic fragmental rocks belong to the Hoyle assemblage or the Bartlett assemblage; assignment to the Hoyle assemblage is based on the interpreted turbiditic origin of these rocks (e.g., Fyon and Crocket 1982).

To the south, across the Porcupine–Destor deformation zone (Figure 11.11), the Hoyle assemblage is juxtaposed against the Bartlett, Eldorado and Bowman assemblages. The northeastern part of the Hoyle assemblage is separated from the Kidd–Munro assemblage by the Pipestone shear zone (Prest 1952, 1953). The nature of the northwestern contact, where metavolcanic rocks of the Kidd–Munro assemblage appear to be interlayered with metasedimentary rocks of the Hoyle assemblage, is not well known.

Figure 11.11. Porcupine–Destor deformation zone near Wawaitin Falls, Thornloe Township. The intense shear fabric is developed within Hoyle assemblage metasedimentary rocks close to the contact with the Bartlett assemblage.

East-striking couplets of metavolcanic and metasedimentary units in the central part of the area (Murphy and Hoyle townships) appear to be bounded by faults (Dunbar 1948) and may represent fault-repeated stratigraphy. The metavolcanic rocks of these couplets are similar to rocks of the Tisdale assemblage. As these metavolcanic-metasedimentary pairs exist near the Hoyle–Tisdale assemblage boundary, this boundary may be tectonic. Locally, along the southeastern boundary, rocks of the Hoyle assemblage are unconformably overlain by "Timiskaming-like" metasedimentary rocks of the Three Nations assemblage (*see* Three Nations and Garrison Assemblages).

From the inferred unconformable relationship between the Kidd–Munro assemblage (*see* Kidd–Munro Assemblage) to the north (inferred to be older) and the Hoyle assemblage to the south (inferred to be younger), rocks of the Hoyle assemblage may be younger than 2714 Ma, the age of the youngest rhyolite in the Kidd–Munro assemblage (Corfu et al. 1989).

Tisdale Assemblage

This greenschist facies assemblage was formerly called the Tisdale group (Dunbar 1948; Pyke 1982). The Tisdale assemblage (TD) is bounded to the south by the Porcupine–Destor deformation zone, which juxtaposes south-facing tholeiitic and komatiitic metavolcanic rocks of the Tisdale assemblage (Figures 11.12 and 11.13) against north-facing, intermediate and felsic calc-alkalic metavolcanic rocks and sulphide facies iron formation of the Deloro assemblage (*see* Deloro Assemblage; Pyke 1982; *see* Figure 11.7). Along the western and northern part of the Tisdale assemblage, metavolcanic flows are interlayered, either structurally and/or stratigraphically, with metasedimentary rocks of the Hoyle assemblage (Dunbar 1948). To the northeast, the Tisdale assemblage is unconformably overlain by metasedimentary rocks of the Timiskaming assemblage; locally, this contact is sheared. The Krist and Porcupine assemblages (*see* Krist Assemblage, Porcupine Assemblage) appear to lie in stratigraphic contact with the Tisdale assemblage (Pyke 1982).

The Tisdale assemblage consists of 2 divisions in the Timmins area: 1) a lower division consisting of ultramafic, pyroxenitic and basaltic komatiites and magnesium-rich tholeiites ("formation" IV, Pyke 1982); and 2) an upper division consisting of iron-rich tholeiitic basalt, interlayered with thin (less than 5 m) carbonaceous metasedimentary rocks and minor rhyolite ("formation" V, Pyke 1982), the later possibly formed by coalesced "felsic" varioles (Comba et al. 1986).

The upper basaltic division was subdivided into the Northern, Central, Vipond and Gold Centre "subgroups" (Graton et al. 1933; Jones 1948; Ferguson et al. 1968). The Northern "subgroup" consists of interlayered massive, pillowed, amygdaloidal basalt flows, flow-top, pillow breccias and interflow carbonaceous sedimentary rocks. The Central "subgroup" is dominated by interlayered massive and amygdaloidal pillowed basalts and thick units of flow breccia. The transition from Northern to Central "subgroups" is marked by the presence of a thick, massive, locally hyaloclastic mafic unit (Ferguson et al. 1968). The Vipond "subgroup" consists of iron-rich, massive, pillowed and variolitic tholeiitic basalt flows, hyaloclastite, and discontinuous lenses of carbonaceous argillite and rare semi-massive pods of pyrite (Ferguson et al. 1968; Rye and Edmonds 1990). The Gold Centre "subgroup", exceeding 150 m in thickness, is dominated by coarse-grained, massive, tholeiitic basalt flows (Ferguson et al. 1968; Rye and Edmonds 1990).

Intruded into the Tisdale assemblage are 2691 to 2688 million-year-old (Corfu et al. 1989; Marmont and Corfu 1989) quartz-feldspar-porphyritic, granodioritic stocks (Pearl Lake, Coniaurum and Gillies Lake porphyries) which coalesce downwards forming a larger intrusion at depth (Rye and Edmonds 1990). Chemically, the porphyries resemble sodic, high-alumina trondhjemites (Burrows and Spooner 1986; Rye and Edmonds 1990). Most porphyries were subjected to intense deformation and alteration, and are now quartz-sericite schists. Minor 2673 million-year-old (Marmont and Corfu 1989) albitite dikes of lamprophyric affinity (Burrows and Spooner 1986) cut the porphyry intrusions (Mason and Melnik 1986).

Figure 11.12. Sheared komatiite which has been subsequently folded by east-west shortening, Deloro Township. Komatiites at southern part of Tisdale assemblage, in contact with Eldorado assemblage. Hammer handle points to the northeast.

Figure 11.13. Polygonally jointed komatiite flow, Tisdale assemblage in northern Tisdale Township. Differential weathering and enhancement of joints is attributed to alteration (D.R. Pyke, Ontario Geological Survey, personal communication, 1978).

Strike and facing directions within the Tisdale assemblage are complex. The metavolcanic units generally dip steeply and strike easterly. Variations in strike and facing can be attributed to the presence of: 1) the northwest-striking Kayorum and South Tisdale synclines (e.g., Ferguson et al. 1968; Pyke 1982); 2) northeast-striking folds, such as the Porcupine, Central Tisdale and North Tisdale synclines (e.g., Ferguson et al. 1968; Pyke 1982); and 3) several east-northeast-striking deformation zones, such as the Porcupine–Destor and Dome shear zones and Hollinger Main faults, some of which mark abrupt facing reversals (e.g., Dome fault, Piroshco and Kettles 1987, 1988). The Dome shear zone is interpreted to represent a branch of the Porcupine–Destor deformation zone (Rogers 1984; Kallio 1990; Hodgson et al. 1990). These deformation zones cut the 2691 to 2688 million-year-old porphyry intrusions (e.g., Marmont and Corfu 1989).

STRUCTURAL EVOLUTION OF THE TISDALE ASSEMBLAGE

The structural evolution of the Timmins area is complex and is summarized below from Piroshco and Kettles (1988), Hodgson (1982) and Hodgson et al. (1990). The Porcupine syncline and the North Tisdale anticline are pre-cleavage folds (e.g., Roberts 1981; Hodgson 1983b) and are interpreted to be the oldest structures. These folds have long axial traces, and therefore may have a shallow plunge. Younger structures include the Gilles Lake shear zone and the Hollinger anticline. The Hollinger anticline may be related to dextral motion along the Gilles Lake shear zone (Hodgson et al. 1990). These younger structures are, in turn, truncated or deformed by the Dome and Porcupine–Destor shear zones (Hodgson 1982; Hodgson et al. 1990).

The oldest recognizable foliation, informally labelled "D1" by Hodgson et al. (1990), strikes west-northwest, dips steeply to the north and is axial planar to the South Tisdale anticline (Piroshco and Kettles 1988; Hodgson et al. 1990). Developed in this foliation plane is a -60° plunging, northeast-trending mineral and dimensional lineation. The youngest foliation, labelled "D2" by Hodgson et al. (1990), is northeast striking and locally is intensely developed defining zones of high strain (Piroshco and Kettles 1988; Hodgson et al. 1990). During this deformation episode, re-activation of the Dome shear zone resulted in a sinistral component of horizontal displacement (Piroshco and Kettles 1988; Hodgson et al. 1990).

GOLD MINERALIZATION IN THE TISDALE ASSEMBLAGE

The Tisdale assemblage is unique in that it hosts most of the lode gold deposits in the Timmins area. Approximately 1780 tonnes of gold have been produced from deposits which occur within or immediately adjacent to the Tisdale assemblage. This gold mineralization occurs in 3 different settings: 1) within mafic metavolcanic rock, in the south-central part of the Tisdale assemblage, close to "D2" foliation (*see* Structural Evolution of the Tisdale Assemblage) regional shear zones and the assemblage boundary (e.g., Hollinger, McIntyre, Coniaurum and Dome mines; e.g., Burrows 1912; Ferguson et al. 1968; Mason and Melnik 1986; Moritz 1988); 2) close to the northern contact with the Hoyle assemblage (e.g., Bell Creek, Marlhill, Owl Creek and Hoyle Pond); and 3) in metavolcanic and metasedimentary rocks of the Tisdale and Three Nations assemblages (*see* Tisdale Assemblage, Three Nations and Garrison Assemblages), respectively, close to the locally deformed, unconformable assemblage contact (e.g., Falconbridge Hoyle, Pamour No.1, Hallnor and Broulan Reef mines; Pyke 1982; Kusic and Olson 1990; West et al. 1990). At least 2 habits of gold mineralization occur—quartz ± carbonate vein (lode-type) and sulphide-rich zones (disseminated-type) which lack quartz veins. In addition to the gold, copper and by-product molybdenite were extracted from a disseminated sulphide and veinlet orebody within the ~2690 million-year-old Pearl Lake porphyry intrusion (Pyke and Middleton 1971; Rye and Edmonds 1990).

Most lode gold-type mineralization occurs in and adjacent to: 1) shear zone-hosted, vertical quartz veins which cut metavolcanic, metasedimentary and intrusive porphyry rocks; and 2) flat veins and "hook quartz veins" localized at closures of small-scale folds, especially within interflow metasedimentary units (Graton et al. 1933; Hall 1985; Rye and Edmonds 1990). Stockwork gold-quartz stringers represent a minor component of vein types which cut sericitic heterolithic intrusion breccias at or near porphyry contacts. Quartz veins are abundant in some of the porphyry intrusions, but auriferous zones are erratic. The quartz (± albite-tourmaline-sericite) veins are locally intensely deformed (Kent 1990) and vary in thickness from centimetres to swarms reaching several metres in thickness (Rye and Edmonds 1990).

In the northeastern part of the Tisdale assemblage, disseminated sulphide replacement-type gold mineralization occurs in the Bell Creek Mine (Kent 1990) where 2 to 10% pyrite and accessory arsenopyrite, pyrrhotite and chalcopyrite predominate, although minor quartz veins are present. In this ore type, roughly 90% of the gold is associated with the disseminated sulphides, which occur within highly altered quartz-carbonate-sericite-sulphide zones.

Regardless of mineralization type, gold occurs as the native element and in fine-grained pyrite, concentrated along sericitic wall rock selvages (Wood et al. 1986). Pyrite is the dominant sulphide, locally making up 10% of the rock. In the Dome Mine, tellurides, including altaite, petzite and tellurobismuthite, occur; however, the tellurides constitute a very minor source of the total gold production. Scheelite is found in minor quantities. Silver is recovered as a by-product of the operation in the ratio of about 1 ounce of silver to 6 ounces of gold for mines in the core of the Porcupine camp to 1 to 10 for deposits which occur farther away. Other sulphides, including pyrrhotite, chalcopyrite, sphalerite, galena and arsenopyrite, rarely occur; when present, chalcopyrite, sphalerite and galena generally correspond to higher gold tenors (Kallio 1990; Labine 1990). Significant amounts of brown tourmaline (dravite) also occur in the veins.

Vein-associated alteration zones generally consist of 3 zones (Fyon and Crocket 1982; Piroshco and Hodgson 1988): 1) an outer zone of chlorite and low-iron dolomite; 2) a central zone of high-iron dolomite to ankerite; and 3) an inner zone of abundant sericite and ankerite. These zones host most of the mineralized vein systems and are commonly coextensive with shear zones. In the northeast, close to the assemblage contact between the Tisdale and Hoyle assemblages, alteration associated with auriferous quartz veins includes the presence of reduced carbon ("grey zones"; Downes et al. 1984). Bright green hydromuscovite is also conspicuous in these veins and is associated with visible gold (Kent 1990).

Although Mason (1986) and Mason and Melnik (1986) proposed a syn-intrusion age for the gold-bearing quartz veins, several lines of evidence are consistent with a younger age. The coincidence of "D2", northeast-striking regional planar and linear fabrics with the orientation of shears, veins, ore shoots and the plunge of the porphyries suggests that alteration and mineralization were coincident with the later stages of regional "D2" deformation (Piroshco and Hodgson 1988; Rye and Edmonds 1990; Hodgson et al. 1990). Auriferous veins cut the 2691 to 2688 million-year-old intrusive quartz-feldspar porphyries (Corfu et al. 1989; Marmont and Corfu 1989). Hydrothermal sericite from altered porphyries has a $^{40}Ar/^{39}Ar$ age of 2633±6 Ma (Masliwec et al. 1986). While this mica age is representative of a thermal event, it cannot be unequivocally related to the precipitation of gold; hence, it can only be stated that gold-quartz veins formed following the emplacement of the felsic porphyritic intrusions.

Krist Assemblage

The greenschist, 2698±4 million-year-old (Corfu et al. 1989), Krist assemblage (KR) appears to lie conformably above the Tisdale assemblage and is overlain by the Porcupine assemblage (Pyke 1982). This assemblage occurs in 2 areas in Tisdale Township; it overlies the western part of the Tisdale assemblage, where part of the contact with the Tisdale assemblage corresponds with the Hollinger Main fault. The second area of Krist assemblage crops out around the Porcupine Syncline (Figure 11.14). Here, a stratigraphic contact with the Tisdale assemblage is inferred (e.g., Ferguson et al. 1968; Pyke 1982). The assemblage consists of calc-alkalic, felsic pyroclastic rock ("Krist" fragmental rocks of Ferguson et al. 1968; "formation" VI of Pyke 1982) associated with carbonaceous metasedimentary units at the top and base.

The 2 geographically separated areas of Krist assemblage rocks are inferred to have been originally continuous, and to have occupied the core of the Porcupine syncline. Later refolding, during the development of the "D1", west-northwest-striking foliation, resulted in erosion and exposure of older underlying rocks of the Tisdale assemblage (Hodgson et al. 1990).

Porcupine Assemblage

The Porcupine assemblage (PP; *see* Figure 11.7) is restricted to the Porcupine Syncline and consists of east- to northeast-striking, steeply dipping, south-facing turbidites. The assemblage conformably overlies 2698±4 million-year-old (Corfu et al. 1989) Krist assemblage. Where folded close to the nose of the Porcupine Syncline, easterly facings occur (Ferguson et al. 1968). The Porcupine assemblage is unconformably overlain by Timiskaming-like metasedimentary rocks of the Three Nations assemblage (*see* Three Nations and Garrison Assemblages). Pyke (1982) argued that the Porcupine, Krist and Tisdale assemblages were approximately time equivalent, since there is an apparent conformable relationship between the 2 assemblages.

Three Nations and Garrison Assemblages

The Three Nations assemblage (TN; *see* Figure 11.7) is a northeast- to east-striking assemblage which extends from the Dome Mine area east to the Nighthawk Lake area. The Three Nations assemblage is bounded to the: 1) northeast by the Hoyle assemblage; 2) northwest and west by the Tisdale assemblage; and 3) south by the Porcupine–Destor deformation zone and Dome shear zone (Burrows 1912; Laird 1931; Ferguson et al. 1968). This assemblage includes the Dome and Three Nations formations of Pyke (1982). The Dome formation of Pyke (1982) consists of interbedded polymictic conglomerate and turbidites. The Three Nations formation of Pyke (1982) consists of interbedded conglomerate and cross-bedded sandstone (Figure 11.15). Units of the Three Nations assemblage are steeply north dipping and south facing (Burrows 1912; Ferguson et al. 1968; Pyke 1982). An unconformable relationship with the stratigraphically underlying Tisdale and Porcupine assemblages is documented at several localities in the Timmins area (Ferguson et al. 1968); although this relationship cannot be confirmed in the Nighthawk Lake area because of poor exposure (Laird 1931). A detrital zircon age of 2679±3 Ma (Corfu et al., in press) indicates that the Three Nations assemblage is at least, in part, younger than the porphyry intrusions of the Timmins area (~2690 Ma, Corfu et al. 1989). If correct, this would imply that the deformed contact between the Preston porphyry and Three Nations assemblage exposed in the Dome Mine area is strictly a fault and/or faulted unconformable contact. The basal conglomerate of the Dome formation is reported to contain clasts of quartz-feldspar porphyry (Pyke 1982, p.44), consistent with the U-Pb zircon ages of the porphyries and Three Nations metasedimentary rocks; however, in many Timiskaming-like rocks, the distinction between porphyry and felsic metavolcanic clasts is difficult to make.

The Garrison assemblage (GR) crops out in Guibord, Michaud and Garrison townships (Satterly 1949a, 1949b; Prest 1952, 1953; Troop 1989), and in Holloway Township (Satterly 1954) as a thin unit of turbidite, arkose, conglomerate and minor iron formation (Satterly 1949a). The contacts of the assemblage are not well understood; however, the metasedimentary rocks are tentatively interpreted to be of Timiskaming-type bounded by shear zones of the Porcupine–Destor deformation zone (Troop 1989).

Lode gold mineralization occurs within the Three Nations assemblage (e.g., Falconbridge Hoyle, Pamour No.1, Hallnor and Broulan mines), adjacent to the unconformity with the metavolcanic rocks of the Tisdale assemblage. This mineralization is described with the Tisdale assemblage.

BELFORD–STRACHAN AREA

Metavolcanic and metasedimentary rocks in the Belford–Strachan area (*see* Figure 11.6) are bounded by 2696 million-year-old (Barrie and Davis 1990) granodiorite intrusions to the north, east and south, and the Kapuskasing Structural Zone to the west (Bennett 1969; Thurston et al. 1977). The belt consists of mafic to felsic metavolcanic rocks and mafic intrusions. Metavolcanic rocks of this belt may continue into the Cochrane–Lake Abitibi area (OGS 1984, 1989). The belt is divided into northern and southern domains by the 2702 million-year-old (Barrie and Naldrett 1990), east-striking Montcalm Gabbroic Complex. The Belford–Strachan area is not subdivided into assemblages.

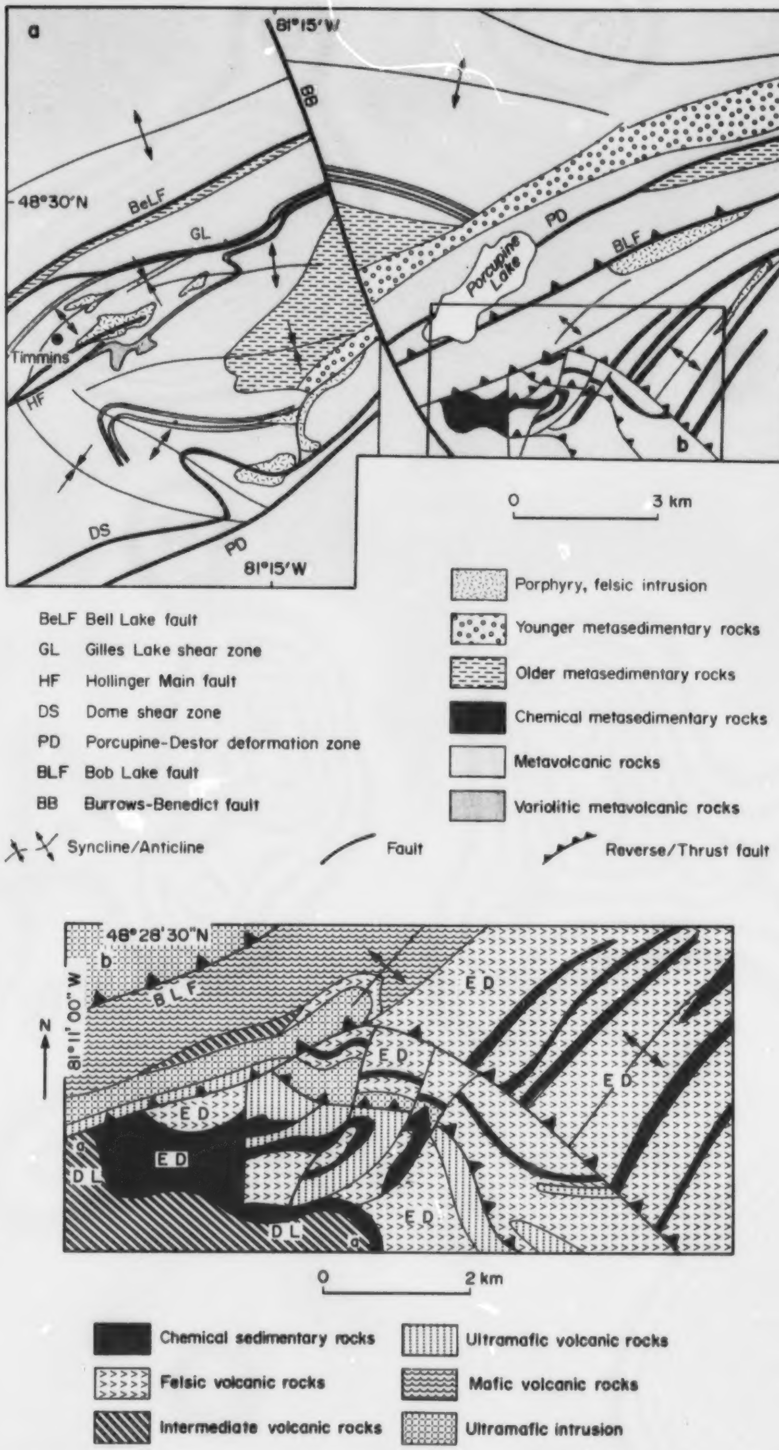

a
81°15'W
BB
48°30'N
BeLF
GL
Timmins
HF
DS
PD
81°15'W
Porcupine Lake
PD
BLF
b
0
3 km
BeLF Bell Lake fault
GL Gilles Lake shear zone
HF Hollinger Main fault
DS Dome shear zone
PD Porcupine-Destor deformation zone
BLF Bob Lake fault
BB Burrows-Benedict fault
Porphyry, felsic intrusion
Younger metasedimentary rocks
Older metasedimentary rocks
Chemical metasedimentary rocks
Metavolcanic rocks
Variolitic metavolcanic rocks
Syncline/Anticline
Fault
Reverse/Thrust fault
48°28'30"N
b
BLF
81°11'00"W
N
ED
ED
ED
ED
ED
ED
DL
DL
a
a
0
2 km
Chemical sedimentary rocks
Felsic volcanic rocks
Intermediate volcanic rocks
Ultramafic volcanic rocks
Mafic volcanic rocks
Ultramafic intrusion

Figure 11.14. Structural geology of the Timmins area. **a)** Generalized geology of the Timmins area, *modified from* Pyke (1982) and Piroshco and Kettles (1987, 1988, 1991). Box outlines region shown in **b** below. **b)** Detail of area outlined in **a** showing mapped and inferred thrust faults in the Eldorado assemblage (ED) and the Deloro assemblage (DL). Thrust fault "a-a" marks the boundary between the Eldorado and Deloro assemblages in this region.

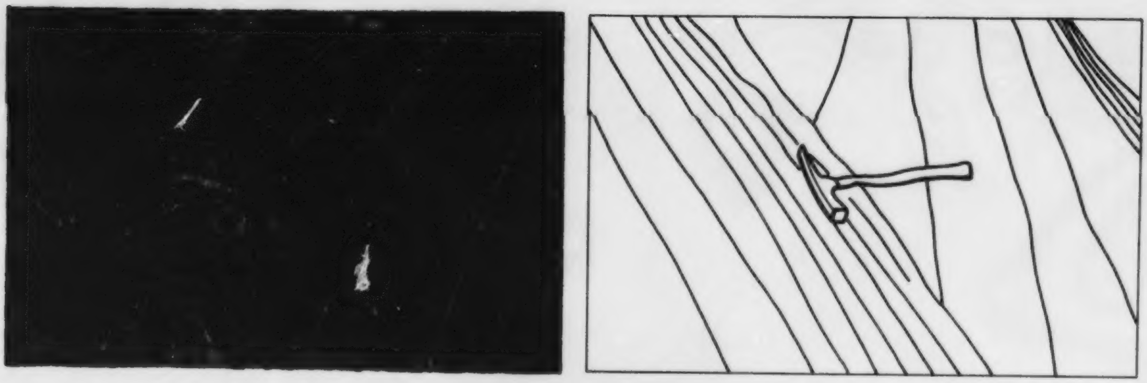


Figure 11.15. Trough cross-bedded quartz arenite in the Three Nations assemblage (Timiskaming-like assemblage) from Shovel outcrop, Whitney Township. The traces of bedding planes are shown on the accompanying sketch.

In the northern part of the belt, tholeiitic and more fractionated basalts showing calc-alkalic affinities are interbedded with east-trending, calc-alkalic, andesitic, dacitic and rhyolitic fragmental rocks (Bennett 1969; Thurston et al. 1977; Barrie 1990a). The felsic metavolcanic rocks consist of, in part, coarse fragmental rocks and porphyritic dacite flows (Bennett 1969).

The Montcalm Gabbroic Complex is a stratiform, layered intrusion, consisting of pyroxenite, gabbro, anorthositic gabbro and ferroan gabbro zones (Barrie and Naldrett 1990). This complex appears to be gradational into overlying metavolcanic rocks. Compositional layering within the Montcalm Gabbroic Complex suggests that the sill faces south (Barrie and Naldrett 1990). Dikes of peridotite, pyroxenite, gabbro and 2700 million-year-old quartz-monzonite-granodiorite cut the Montcalm Gabbroic Complex (Barrie and Naldrett 1990).

In the southeastern part of the belt, mafic and intermediate metavolcanic rocks predominate and are interlayered with fine-grained metasedimentary rocks (Bennett 1969; Thurston et al. 1977).

Amphibolite facies metamorphic conditions were attained adjacent to gneissic felsic intrusions (Thurston et al. 1977). Almandine-amphibolite facies conditions were attained adjacent to the Kapuskasing Structural Zone (Bennett 1969). Farther to the east, greenschist facies conditions prevailed (Bennett 1969).

Moderate (less than 50°) northwest-dipping foliation and bedding characterize the western part of the area, possibly reflecting the influence of the Ivanhoe Lake cataclastic zone (Bennett 1969). East of the Ivanhoe Lake cataclastic zone, penetrative foliations generally strike east to northeast (Bennett 1969). Along the southern margin of the belt, metamorphic layering in amphibolitized basalt strikes east (Bennett 1969). Along the eastern contact with the granodiorite intrusions, a northwest-dipping (greater than 40°), southwest-striking shear zone is present. S-shaped minor folds plunge greater than 60° to the southwest, suggesting an oblique, sinistral (normal; granite-side-up) sense of slip (Bennett 1969).

The Montcalm Gabbroic Complex contains a magmatic nickel-copper deposit (3.56 million tonnes, 1.44% Ni, 0.68% Cu; Barrie and Naldrett 1990), which is localized within clinopyroxene cumulates of the gabbro zone.

ASSEMBLAGES IN THE WATABEAG AREA

Deloro Assemblage

The Deloro assemblage (DL; *see* Figure 11.7; middle "formation" of Deloro group, Pyke 1982) is exposed in the Shaw Dome, southeast of Timmins. This assemblage consists predominantly of pillowed, commonly amygdaloidal and plagioclase-phyric, calc-alkalic basalt and andesite (Pyke 1982; Figure 11.16). The assemblage is characterized by a regional aeromagnetic low with superimposed aeromagnetic highs that coincide with ultramafic-mafic intrusions. The plagioclase-rich differentiate of a dunite intruding the Deloro assemblage is 2707±3 Ma (Corfu et al. 1989), providing a minimum age for the assemblage.

In general, the units within the Shaw Dome are outward facing with moderate to shallow outward dips. Foliations within the Shaw Dome also dip moderately outward (Pyke 1982). The Shaw Dome may be an anticlinal, west-northwest-trending structure related to a series of anticlines and synclines west of the Kenogamissi Batholith (*see* Figure 11.9).

In southwestern Whitney Township, the boundary between the Deloro assemblage and the Eldorado assemblage (felsic and ultramafic metavolcanic rocks with iron formation; *see* Eldorado and Bartlett Assemblages) is marked by well-developed schistosity and a marked contrast in structural style on either side of the boundary (Piroshco and Kettles 1987, 1988, 1991). In this area, the contact appears to be a low-angle fault (thrust?) related to folds and thrusts in the structurally higher Eldorado assemblage (Piroshco and Kettles 1987, 1988, 1991). The remainder of the contact between the Deloro and Eldorado assemblages in Eldorado, Langmuir and Carmen townships is not as well understood because of poor exposure and limited structural analysis. Although a low-angle fault zone may extend around the

Figure 11.16. Pillowed calc-alkalic andesite containing silica-filled vesicles from Deloro assemblage, Deloro Township. Note the large size of pillows.

Shaw Dome (*see* Eldorado and Bartlett Assemblages), Green (1978) suggested that the contact between the upper units of the Deloro assemblage and the overlying units of the Eldorado assemblage is, at least in part, a stratigraphic contact.

Deposits of asbestos and magnesite are associated with some of the serpentinized and carbonatized ultramafic rocks intruding the Deloro assemblage (Pyke 1982; Kretschmar and Kretschmar 1986). Minor gold was extracted from quartz-carbonate veins at the Faymar Mine (Tremblay 1946). These auriferous veins cut calc-alkalic andesite flows and quartz-feldspar porphyry dikes; little alteration is present in the wall rocks of the veins (Fyon et al. 1983).

Eldorado and Bartlett Assemblages

The Eldorado assemblage (ED; *see* Figure 11.7) consists of intermediate to felsic metavolcanic rocks, iron formation and associated metasedimentary rocks and ultramafic flows and intrusions (Pyke 1982). The assemblage crops out around the Shaw Dome and may correlate with the Bartlett assemblage (BL) east of the Kenogamissi Batholith (*see* Figure 11.7). The Eldorado assemblage has a complex aeromagnetic pattern reflecting its lithologic and structural complexity. The Bartlett assemblage is similar to the Eldorado assemblage in that it contains a significant component of iron formation and felsic fragmental rocks (Pyke 1978), but differs in that ultramafic flows are apparently lacking.

Felsic metavolcanic rocks of the Eldorado assemblage are dominated by fragmental varieties ranging from dacitic to rhyolitic tuffaceous rocks that grade in their crystal and lithic proportions and clast size to breccias that are commonly poorly to unsorted and poorly to nongraded (Pyke 1982). Iron formation in these assemblages consists predominantly of oxide and sulphide facies (Pyke 1982) and is associated with minor chert, fine-grained sandstone, siltstone and shale (Pyke 1982). Although ultramafic flows are included in the Eldorado assemblage, these flows may represent the base of the Bowman assemblage that has been folded into the top of the Deloro assemblage.

In southern Tisdale and Whitney townships, the base of the Eldorado assemblage appears to have been thrust over the Deloro assemblage (Piroshco and Kettles 1987, 1988, 1991). The Eldorado assemblage in this area has been internally imbricated by south-verging thrusts and highly folded (Piroshco and Kettles 1987, 1988, 1991). Around the southern margin of the Shaw Dome, the Eldorado assemblage locally exhibits well-developed schistosity that generally dips shallowly to moderately away from the Shaw Dome (Pyke 1982). These shallowly dipping fabrics may correlate with the thrust faults and related fabrics to the northeast (Piroshco and Kettles 1987, 1988, 1991). Hence, a low-angle fault at the base of the Eldorado assemblage may extend around the Shaw Dome.

A hornblende-bearing gabbroic rock, interpreted to be a dunitic differentiate which cuts felsic metavolcanic rocks of the Deloro assemblage (Pyke 1982), has an age of 2707±3 Ma (Corfu et al. 1989). This intrusion has been interpreted to be genetically related to the komatiitic flows which crop out within the Eldorado assemblage (Pyke 1982). Consequently, the 2707 Ma age would also represent the maximum age of the komatiitic and interlayered felsic metavolcanic flows within the Eldorado assemblage. This would negate correlation of the Eldorado assemblage with the 2727 million-year-old (Corfu et al. 1989) Bartlett assemblage. However, the genetic relationship between the ~2707 million-year-old ultramafic intrusion, which cuts the Deloro assemblage, and the ultramafic metavolcanic and intrusions of the Eldorado assemblage, is equivocal, based on geochemical criteria (A.H. Green, Falconbridge Limited, personal communication, 1990). Consequently, the Eldorado assemblage is not necessarily 2707 Ma.

Alternatively, the felsic metavolcanic rocks and iron formation of the Eldorado and Bartlett assemblages may be correlated, as suggested by their similar electromagnetic and geochemical patterns (Pyke 1982; MERQ–OGS 1983). East of the Kenogamissi Batholith, a felsic member of the Bartlett assemblage has been dated at 2727±1.5 Ma (Corfu et al. 1989). If the metavolcanic rocks of the Bartlett and Eldorado assemblages are correlative, and thus are ~2727 Ma, then the ~2707 million-year-old ultramafic intrusion could not represent the age of komatiitic rocks at the top of the Eldorado assemblage.

Mineralization within the Eldorado assemblage includes komatiite-associated, nickel-copper sulphide mineralization associated with ultramafic flows (Pyke 1982; Green and MacEachern 1990). The nickel sulphide mineralization occurs along the southern side of the Shaw Dome, where 5 nickel-sulphide deposits and several showings (e.g., Redstone, Hart (Tontine), McWatters, Langmuir No.1 and Langmuir No.2) occur within the basal komatiitic flows. The ultramafic flows, which host the nickel-sulphide deposits, may define a stratigraphic horizon overlying the felsic to intermediate metavolcanic rocks of the Eldorado assemblage (Pyke 1970; Green and MacEachern 1990). The horizon is associated with the thickest accumulation of sulphidic metasedimentary rocks (Green 1978).

Within the Bartlett assemblage, komatiite-associated nickel sulphide mineralization occurs at the Texmont Mine, which is located at the boundary of Bartlett and Geikie townships. Like mineralization in the Eldorado assemblage, nickel-sulphide mineralization in the Bartlett assemblage occurs in an ultramafic komatiite flow near the upper part of a felsic metavolcanic unit which is intercalated with iron formation (sulphidic interflow metasedimentary units). The metavolcanic unit, which hosts the Texmont Mine, may be stratigraphically correlative with the metavolcanic unit which hosts the deposits in the Langmuir–Eldorado area.

A second resource, magnesium in magnesite, is associated with a carbonatized ultramafic intrusion (Pyke 1982; Kretschmar and Kretschmar 1986).

Peterlong Assemblage

The Peterlong assemblage (P; *see* Figure 11.7; Pyke 1978) occurs east of the Kenogamissi Batholith and is a homoclinal east-facing succession of predominantly massive and pillowed flows of probable tholeiitic affinity. Minor felsic metavolcanic and metasedimentary rocks are associated with the assemblage. The assemblage is bound to the east by the Muskasenda gabbro and the Bartlett assemblage. The Bartlett–Peterlong assemblage contact either is not exposed or is intruded by the Muskasenda gabbro (Pyke 1978, Map 2345). The low and flat aeromagnetic pattern of this assemblage is consistent with the Peterlong assemblage being a dominantly magnesium-rich tholeiitic metavolcanic assemblage. Foliations within the Peterlong assemblage dip steeply to moderately away from the Kenogamissi Batholith (Pyke 1978).

The Peterlong assemblage was previously correlated with the Deloro assemblage (MERQ–OGS 1983). Jensen and Langford (1985), however, separated the 2 assemblages (Jensen and Langford 1985, Figure 5). The probable tholeiitic nature of the Peterlong assemblage (Pyke 1978) versus the calc-alkalic nature of the Deloro assemblage (Pyke 1982, Figure 20a) supports the Jensen and Langford (1985) interpretation.

The age of the Peterlong assemblage is unknown. However, in the southern Abitibi greenstone belt, the oldest units are exposed adjacent to large batholiths, for example: 1) the ~2730 million-year-old Hunter Mine adjacent to the Lake Abitibi Batholith; 2) the ~2747 million-year-old unit in the Catharine–Pacaud assemblage adjacent to the Round Lake batholith; and 3) the ~2727 million-year-old Bartlett assemblage adjacent to the Kenogamissi Batholith. Therefore, an interpreted "old" age (greater than 2720 Ma) is favoured for the Peterlong assemblage.

Bowman Assemblage

The Bowman assemblage (BM; Goodwin 1977) outcrops south and east of the Shaw Dome and east of the Kenogamissi Batholith (*see* Figure 11.7). The assemblage extends westward from the southern side of the Porcupine–Destor deformation zone near Night Hawk Lake, around the southern margin of the Shaw Dome, to the northern side of the Adams Pluton. It reappears on the southeastern side of the Adams Pluton and extends southwards into Semple and Hutt townships. The assemblage generally faces toward the region between the Porcupine–Destor deformation zone and Larder–Cadillac shear zone. This assemblage includes peridotitic and basaltic komatiites interlayered with komatiites and magnesium-rich basalt (Pyke 1982; Leahy 1965).

In the vicinity of the Shaw Dome, the Bowman assemblage has an aeromagnetically high signature, likely reflecting a high component of variably serpentinized ultramafic rocks. Eastward, towards Matheson, the assemblage has an aeromagnetically low signature, possibly reflecting a greater component of magnesium(?)-rich tholeiitic basalt units.

Near the southern Bowman–Geikie assemblage boundary, moderate to steep, north-dipping schistosity is present. Preliminary work (B. Hrabi, Queen's University, Kingston, personal communication, 1990) indicates that, at least locally, a shallow, north-dipping shear zone occurs near the boundary between the assemblages and correlates with the shallow to moderately dipping schistosity delineated by Bright (1984). Near the western boundary of the same 2 assemblages, south of the Geikie Pluton, schistosity strikes northeast and dips steeply north. The schistosity trends of Bright (1984) indicate a period of folding postdating the development of the schistosity. Although possible, there are insufficient data to identify the boundary between the 2 assemblages as a folded shear zone.

The Porcupine–Destor deformation zone separates the Bowman assemblage from the Hoyle assemblage to the north. Near the Shaw Dome, a contact between an ultramafic flow of the Bowman assemblage and a sulphide facies iron formation of the Eldorado assemblage is conformable (Pyke 1982). However, the Eldorado assemblage itself is highly folded and sheared (Pyke 1982; Piroshco and Kettles 1987, 1988, 1991), reducing the confidence of interpretations of stratigraphic relationships between the 2 assemblages. Ultramafic-mafic intrusions in the Deloro and Eldorado assemblages may be feeders to the ultramafic flows of the Bowman assemblage (Pyke 1982). One possible interpretation is that Bowman assemblage rocks were deposited on the Eldorado assemblage and its equivalents and were then subsequently displaced along a low-angle shear zone located at the original interface between the 2 assemblages.

On the eastern side of the Kenogamissi Batholith, the contact between the Bartlett and Bowman assemblages is not exposed (Pyke 1978). The boundary between the 2 assemblages does, however, coincide approximately with an abrupt change in structural style between north-trending, east-facing units of the Peterlong and Bartlett assemblages and the east-trending, highly folded Geikie assemblage. In addition, a fault extending north-northeast from the Kenogamissi Batholith separates the different assemblages and may account for the contrast in structural style between the assemblages (*see* Figure 11.9).

West of Night Hawk Lake, a 2705±10 million-year-old (Corfu et al. 1989) granodiorite intrudes the Bowman assemblage. A ~2704 million-year-old (Corfu et al., in press) felsic metavolcanic unit at the interface between the Bowman and Watabeag assemblages may define the top of the Bowman assemblage. In addition, a 2707±3 million-year-old (Corfu et al., in press) dunite sill may be a feeder to ultramafic rocks of the Bowman assemblage (Pyke 1982; *see* Eldorado and Bartlett Assemblages). These data may indicate that the Bowman assemblage units are ~2705 Ma. This age is significantly younger than the ~2715 million-year-old ultramafic flow-bearing Kidd–Munro assemblage, but is comparable to the ~2705 million-year-old ultramafic flow-bearing Larder Lake assemblage.

Geikie Assemblage

The Geikie assemblage (GK; *see* Figure 11.7) crops out predominantly to the north, east and south of the Geikie Pluton (Pyke 1978). The bulk of this assemblage consists of tholeiitic, pillowed to massive, mafic metavolcanic rocks and minor intermediate to felsic metavolcanic rocks (Pyke 1978). The Geikie assemblage displays well-defined aeromagnetic high- and low-intensity domains which may reflect iron- and magnesium-rich units. Locally, the basalt is variolitic (Pyke 1978; Bright 1984). A felsic metavolcanic unit north of the Geikie Pluton has been dated at 2703±1.5 Ma (Corfu et al. 1989). To the west and south, the Geikie assemblage is in contact with the Bowman and Eldorado assemblages (*see* Bowman Assemblage, Eldorado and Bartlett Assemblages for description of contacts). To the east, most of the Geikie assemblage is unconformably overlain by Paleoproterozoic rocks of the Huronian Supergroup. The remainder is in contact with tholeiitic rocks, although the nature of the later contact is not known.

The Geikie assemblage is folded about east-striking axial surfaces, with 3 major folds being recognized: an anticline cored by the Geikie Pluton and synclines north and south of this pluton. In Zavitz Township, there are numerous facing reversals (Bright 1984) indicating the presence of second order folds related to the southern syncline.

Watabeag Assemblage

The Watabeag assemblage (WA; *see* Figure 11.7) crops out west of the Watabeag Batholith, and extends from the Bowman assemblage, south of the Porcupine–Destor deformation zone, to the Larder–Cadillac shear zone (*see* Figure 11.9). Little is known of this assemblage because it is poorly exposed. The assemblage yields a flat aeromagnetic pattern comparable to the patterns of the Blake River, Halliday and Skead assemblages, but is distinct from the banded pattern of the Kinojevis North assemblage and the convoluted pattern of the Kinojevis South assemblage (Figure 11.17). The age of this assemblage is unknown.

With the exception of the Timiskaming assemblage and the older assemblages adjacent to the Kenogamissi Batholith, the dated units between the Porcupine–Destor deformation zone and the Larder–Cadillac shear zone are between 2705 and 2698 Ma (Corfu et al. 1989; F. Corfu, Jack Satterly Geochronology Lab, Royal Ontario Museum, personal communication, 1990). Therefore, the Watabeag assemblage may also be between 2705 and 2698 Ma.

In Fallon and Fasken townships, Pyke (1973) delineated a large component of felsic metavolcanic rock within the Watabeag assemblage. Rocks of the Watabeag assemblage that crop out just north of the Larder–Cadillac shear zone have been correlated with the calc-alkalic Blake River assemblage (MERQ–OGS 1983). South of the Porcupine–Destor deformation zone, Pyke (1970, 1976) identified mafic metavolcanic rocks and felsic fragmental units. Subsequent drilling has revealed that many rocks in this area are andesitic to dacitic fragmental metavolcanic rocks (OGS 1986a–d). All of the above observations, combined with the flat and low aeromagnetic pattern, suggest that the Watabeag assemblage consists predominantly of intermediate to felsic metavolcanic rocks, not tholeiitic basalt as was previously thought (e.g., MERQ–OGS 1983).

Near the contact between the Watabeag and Bowman assemblages is a felsic metavolcanic unit containing east-striking schistosity (Leahy 1965; Pyke et al. 1973). The schistosity strikes towards, and is along strike from, an east-trending aeromagnetic discontinuity. This discontinuity is interpreted to be a fault, since it marks the termination of some major metavolcanic and metasedimentary assemblages (e.g., Hoyle assemblage). Elsewhere, the Watabeag–Bowman contact is not exposed. The eastern margin of the Watabeag assemblage is in contact with the Kinojevis North and Kinojevis South assemblages along north-trending faults that are spatially associated with a high density of north-striking Matachewan dike swarms. The western Watabeag assemblage is unconformably overlain by the Paleoproterozoic Huronian Supergroup. The relationships between the Watabeag assemblage and assemblages to the south are unknown.

Halliday Assemblage

The Halliday assemblage (HL; *see* Figure 11.7) consists of a central region of rhyolitic to dacitic flows, breccias and tuffs, and an outer region of andesitic to basaltic flow and pyroclastic rocks (Bright 1970). Graphitic tuff and slate are reported from the northern Halliday assemblage. Iron formation is not an important constituent of this assemblage, in contrast to the Eldorado and Bartlett assemblages to the north.

The aeromagnetic pattern of the assemblage is flat, low in intensity and peppered along its southern flank with local aeromagnetic highs which correlate with gabbroic to peridotitic intrusions. The flat and low aeromagnetic intensity contrasts with the more variable and higher-intensity aeromagnetic pattern of other felsic metavolcanic assemblages to the north (e.g., Eldorado assemblage), but is similar to the Skead assemblage east of the Round Lake batholith and the Blake River assemblage (*see* Figure 11.17).

The age of this assemblage is not known. Its similarity with the Skead, Blake River and Gauthier assemblages suggests it could be about 2700 Ma, although its spatial association with the Bartlett assemblage suggests it could be ~2730 Ma.

In general, fabrics within the Halliday assemblage are steeply dipping and northeast trending. Facing directions are limited in number and no pattern is recognizable. Bright (1970) referred to this assemblage of largely felsic metavolcanic rocks as a dome; however, there is little structural or younging data to support such an interpretation.

Northeast-striking, steeply dipping shear zones cut the Halliday assemblage and continue into the younger, Neoarchean metasedimentary rocks (Midlothian assemblage) along its eastern flank (Bright 1970). These shear zones do not appear to displace the contact between felsic metavolcanic rocks of the Halliday assemblage and the metasedimentary rocks.

The assemblage is bounded to the: 1) south by the unconformably overlying Paleoproterozoic Huronian Supergroup; 2) east by a deformed, disconformable, younger assemblage of Neoarchean metasedimentary rocks (Midlothian assemblage) and unconformably overlying Huronian Supergroup; and 3) west and north by ultramafic-mafic rocks of the Bowman assemblage. The nature of the Halliday–Bowman assemblages boundary is not understood. Bright (1970) interpreted rocks of the Halliday assemblage to be interstratified with the mafic metavolcanic rocks of the Bowman assemblage. R. Shegelski (Esso Minerals, personal communication, 1989) mapped a distinctive coarse clastic rock containing predominantly ultramafic and felsic clasts

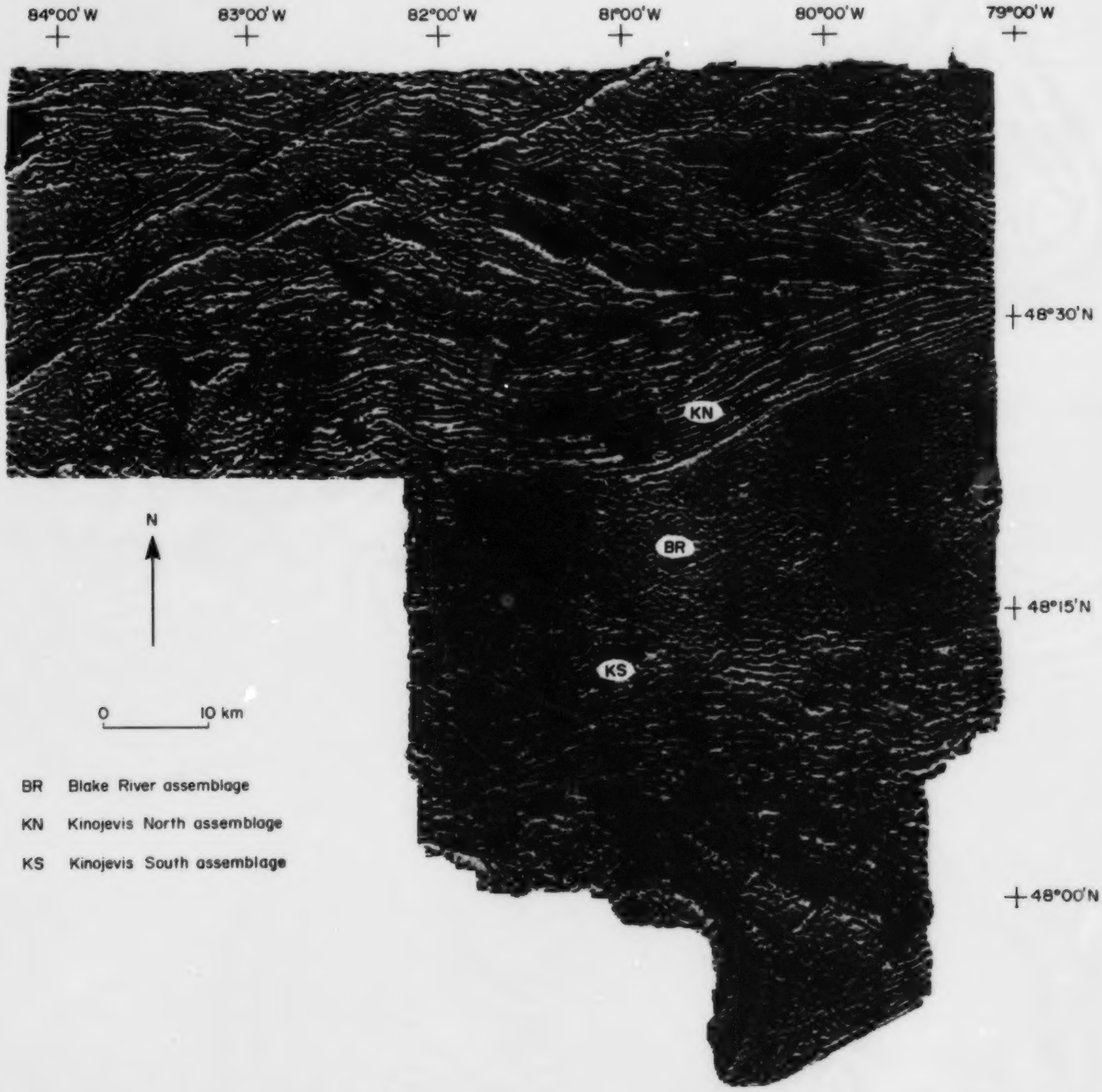

84°00'W
83°00'W
82°00'W
81°00'W
80°00'W
79°00'W
48°30'N
48°15'N
48°00'N
KN
BR
KS
N
0
10 km
BR Blake River assemblage
KN Kinojevis North assemblage
KS Kinojevis South assemblage

Figure 11.17. Map of the second vertical derivative of the total magnetic field centred on the Blake River assemblage in Ontario. Note that: 1) the Blake River assemblage is characterized by a flat aeromagnetic pattern, the Kinojevis North assemblage by a banded aeromagnetic pattern, and the Kinojevis South assemblage by a convoluted aeromagnetic pattern; and 2) that the banding characterizing the Kinojevis North assemblage does not wrap around the apparent fold closure at the western end of the Blake River assemblage, suggesting the presence of a structural discontinuity.

interlayered with graphitic pyritic shale at the boundary between the 2 assemblages with no apparent interstratification of members of the 2 assemblages.

The Stairs gold mine (past producer) is situated at the boundary between the Halliday assemblage and the Midlothian assemblage of metasedimentary rocks (Bright 1970). The mineralized zone is associated with silicification, sericitization, carbonatization and shearing.

Kinojevis North Assemblage

The Kinojevis North assemblage (KN; *see* Figure 11.7) is a steeply dipping, south-facing succession of pillowed, tholeiitic basalt and minor rhyolite (Jensen and Langford 1985). Interflow metasedimentary rocks, including chert, carbonaceous siltstone, lithic wacke and crystal tuff, are scarce. A metarhyolite member of this assemblage has been dated at ~2701 Ma (F. Corfu, Jack Satterly Geochronology Lab, Royal Ontario Museum, personal communication, 1990). Metabasalt members, laterally continuous over tens of kilometres, form distinct magnesium- and iron-rich units. Consequently, the aeromagnetic pattern of this assemblage (*see* Figure 11.17) displays alternating bands of high and low intensity (Lestros et al. 1983). Some flows are locally feldspar-phyric and/or variolitic (Jensen and Langford 1985; Fowler et al. 1987).

The Kinojevis North assemblage tholeiitic basalt and interbedded rhyolite members are characterized by either slightly depleted, flat, or slightly enriched LREE patterns with either no, or slight, negative europium anomaly (Smith 1980; Capdevila et al. 1982; Fowler and Jensen 1989). The lack of LREE enrichment in the rhyolite members indicates that assimilation of evolved, large-ion-lithophile (LIL)-enriched crust is not a viable mechanism for the generation of the Kinojevis North assemblage (Fowler and Jensen 1989). Partial melting of variably depleted to undepleted mantle peridotite has been cited as a means to account for the REE patterns (Smith 1980; Capdevila et al. 1982; Fowler and Jensen 1989).

The Kinojevis North assemblage is truncated to the north by the Porcupine–Destor deformation zone, to the west by the Watabeag Batholith and north-trending faults, and to the south it is in contact with the Blake River assemblage (*see* Blake River Assemblage for discussion of contact relationships).

The Ross Mine, situated south of the Porcupine–Destor deformation zone near Holtyre, produced ~30 tonnes of gold and 46.6 tonnes of silver (Meyer et al. 1989). The setting of this mine is currently under investigation by D.G. Troop (Ontario Geological Survey). Mineralization is situated in a zone of high strain and intense alteration (Troop 1985, 1986) and is spatially associated with alkali basalts, syenitic intrusions, greywackes and conglomerates of possible Timiskaming-type affinity and tholeiitic metabasalts (Troop 1989, 1990). The regional setting of the mine, then, is within a high-strain zone associated with a package of Timiskaming-like rocks, that occurs near the boundary between tholeiitic rocks to the south (Kinojevis North assemblage) and tholeiitic plus komatiitic rocks to the north (Kidd–Munro assemblage) (D.G. Troop, Ontario Geological Survey, personal communication, 1991).

Two other gold mines and several gold occurrences are located in the northern part of the Kinojevis North assemblage, along a deformation zone. The deformation zone is characterized by both ductile and brittle deformation textures and strikes approximately 070. The relationship between the 070-striking structure and the Porcupine–Destor deformation zone is not known. This mineralized zone lies to the south of the Porcupine–Destor deformation zone. The Holt–McDermott and Canamax Resources East Zone are the 2 principal gold mines along this structure. Reserves at the Holt–McDermott mine are approximately 3628 tonnes at 5.21 grams per tonne Au (Meyer et al. 1989).

Kinojevis South Assemblage

The main units of the Kinojevis South assemblage (KS; *see* Figure 11.7) include massive and pillowed metabasalt of tholeiitic affinity (e.g., Goodwin 1979). Only very local, and minor, metamorphosed interbedded iron formation, tuff and clastic metasedimentary rocks are present (Thomson 1943; Thomson and Griffis 1944). Metabasalt members locally display amygdules and/or spherules. Intrusions include gabbro, diorite, quartz gabbro, quartz diorite, granite, syenite and feldspar ± quartz porphyry. The Kinojevis South assemblage is very similar to the Kinojevis North assemblage (Jensen 1985a; Jensen and Langford 1985). Furthermore, Jensen (1989) mapped some units of the Kinojevis South assemblage around the western margin of the Blake River assemblage; how these units link with Kinojevis North assemblage units is not clear. It seems likely, however, that the Kinojevis North and Kinojevis South assemblages are related and that the ~2701 Ma age for the metarhyolite from the Kinojevis North assemblage (F. Corfu, Jack Satterly Geochronology Lab, Royal Ontario Museum, personal communication, 1990) also represents the approximate age of the Kinojevis South assemblage. A syenitic phase of the Winnie Stock, the main felsic intrusion within the Kinojevis South assemblage is 2677±2 Ma (Frarey and Krogh 1986).

The assemblage is folded, faulted, and intruded by numerous small felsic intrusions and consequently displays a much more convoluted aeromagnetic pattern than the Kinojevis North assemblage (*see* Figure 11.14). The major structures include the Beaver Lake Syncline and the Spectacle Lakes Anticline, the latter being cored by the calc-alkalic Gauthier assemblage (Thomson and Griffis 1944). These structures are upright, east trending, and shallowly to moderately plunging (Hamilton 1986). The Spectacle Lake Anticline has Timiskaming assemblage rocks on both limbs (Thomson and Griffis 1944; Jackson 1988), and therefore, is a post–Timiskaming structure. The axial surface of the anticline is truncated by the Larder–Cadillac shear zone, and therefore, predates the final development of this shear zone.

Metavolcanic rocks of the Kinojevis South assemblage are bounded to the: 1) north by the calc-alkalic Blake River assemblage (*see* Blake River Assemblage); 2) south by the Timiskaming assemblage; and 3) west by the Proterozoic Huronian Supergroup and the Archean Watabeag Batholith.

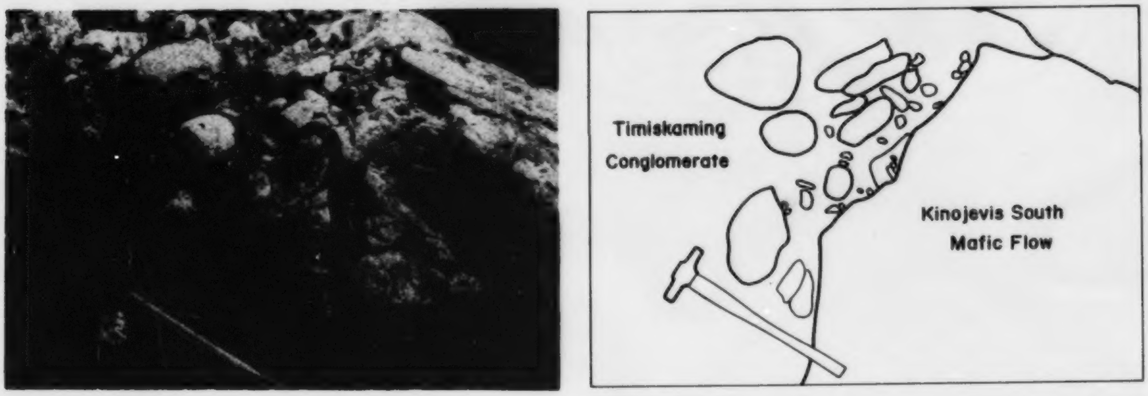


Figure 11.18. Timiskaming assemblage conglomerate (~2685 to 2675 Ma) in depositional contact with Kinojevis South assemblage (~2700 Ma) massive mafic flow, McVittie Township. Accompanying sketch delineates contact.

The Kinojevis South assemblage is unconformably overlain by the ~2685 to 2675 million-year-old (Corfu et al., in press) Timiskaming assemblage to the south (Figure 11.18). Detailed mapping indicates that the Kinojevis South assemblage was folded prior to the deposition of the Timiskaming assemblage (Thomson 1946).

The Kinojevis South assemblage is more highly folded and contains more intrusions (e.g., Winnie Stock and Crosby gabbro sill; Figure 11.19) compared to the Kinojevis North assemblage, which is a steeply dipping, homoclinal, south-facing succession. Therefore, the 2 assemblages appear to have experienced different structural and magmatic histories.

Significant mineralization is known from the Kinojevis South assemblage; however, there is some gold mineralization associated with a shear zone near the Kinojevis South–Gauthier assemblages boundary (Thomson and Griffis 1944).

Blake River Assemblage

The Blake River assemblage (BR; *see* Figure 11.7) is underlain by folded and faulted, mafic to felsic calc-alkalic metavolcanic rocks and small mafic and felsic intrusions (e.g., Jensen 1972, 1975a, 1975b; Hogg 1964; Jensen and Langford 1985). A rhyolite from the centre of the Blake River assemblage is ~2701±2 Ma (Corfu et al. 1989). The assemblage has a regional low and flat aeromagnetic pattern (*see* Figure 11.17) which is similar to the aeromagnetic pattern of the Skead and Halliday assemblages. From north to south, there is a general decrease in total aeromagnetic intensity, which may indicate a fundamental subsurface asymmetry to the Blake River assemblage or its substrate.

Metavolcanic rocks of the Blake River assemblage range from basalt through to rhyolite, with basalt and andesite being dominant. Dacite and rhyolite are abundant in the centre of the assemblage (Jensen 1985a; Jensen and Langford 1985). Small intrusions include gabbro, quartz gabbro, diorite and quartz diorite. Metasedimentary rocks such as iron formation and interflow turbidites are very rare.

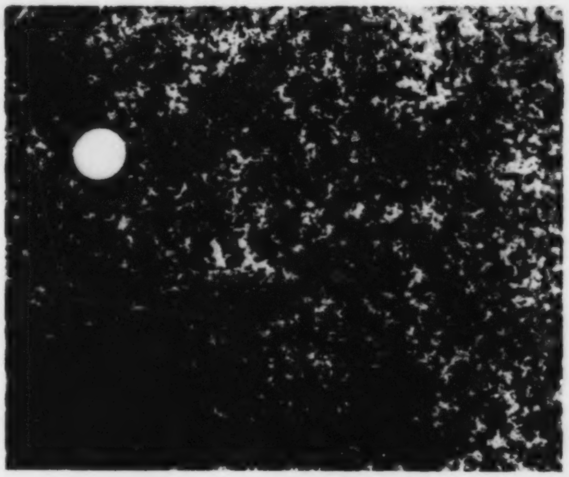


Figure 11.19. Coarse-grained Crosby Lake hornblende-bearing gabbro, McVittie Township.

Units of the Blake River assemblage are shallow to moderately dipping. In general, units at the margin of the assemblage face towards the centre of the assemblage, suggesting a synclinorium. In addition, numerous facing reversals delineate several regional east-trending folds. The centre of the assemblage is occupied by an anticlinal structure cored by felsic intrusions. This structure may in part reflect the presence of an original volcanic centre (e.g., Dimroth et al. 1982). Some facing reversals correlate with mapped faults (e.g., Mulven fault; Hogg 1964; Jensen 1975a) and although not apparently prominent regional features, seismic reflection data indicate that the Mulven and/or Misema–Mist faults penetrate to a depth of at least ~15 km below the surface of the Earth (Jackson, Sutcliffe et al. 1990).

In the eastern part of the assemblage, northeast-striking faults are well developed, and in the western part of the area, northwest-striking faults are prevalent. The northwest-striking set may be related to Timiskaming Rift faults that extend northwest from Lake Timiskaming (Kumarapeli and Saull 1966). The significance of the northeast-striking faults is uncertain; however, tectonic events that have developed approximately northeast-trending structures include northwest-directed thrusting of the Grenville Province and the intracratonic thrusting in the Kapuskasing Structural Zone (Percival and Card 1985).

Units of the Blake River assemblage, near the boundary with the Kinojevis North assemblage, are subparallel. This subparallelism has been interpreted as evidence of stratigraphic continuity across the Kinojevis North assemblage and Blake River assemblage boundary (Jensen and Langford 1985; Jensen 1989). Two lines of evidence suggest that this may not be the case. First, the internal structure of the Blake River assemblage is characterized by shallow to moderate dips and numerous east-trending folds, whereas the Kinojevis North assemblage is a steeply dipping, south-facing, homoclinal unit. Secondly, a map of the second vertical derivative of total field aeromagnetic intensity (*see* Figure 11.17) shows that units of the Kinojevis North assemblage maintain an easterly orientation at the western closure of the Blake River assemblage and do not, in general, curve around the closure (*see* Figure 11.17). These observations suggest that the contact may be tectonic, however, an original stratigraphic contact is possible.

The contact between the Blake River assemblage and the Kinojevis South assemblage is not well mapped; however, aeromagnetic maps indicate a sharp break between the "flat" field of the Blake River assemblage and the highly variable field of the Kinojevis South assemblage (*see* Figure 11.17). In southern Bernhardt and Morrissette townships, pillowed mafic metavolcanic rocks (presumably Kinojevis South assemblage) strike north and appear to terminate against east-southeast-striking intermediate metavolcanic rocks (presumably Blake River assemblage) (Rupert and Lovell 1970). The region between the contrasting trends displays intense folding, shearing, carbonate alteration, and also contains Timiskaming-like sedimentary rocks (Rupert and Lovell 1970). These observations may indicate that the boundary between the Blake River and Kinojevis South assemblages is now a structural contact; however, an original stratigraphic contact cannot be ruled out.

Most of the metavolcanic rocks of the Blake River assemblage in Ontario are part of a calc-alkalic suite (Goodwin 1979; Smith 1980; Jensen and Langford 1985) and are characterized by LREE enrichment and prominent negative europium anomalies in dacitic to rhyolitic rocks (Smith 1980; Fowler and Jensen 1989). There is no consensus on the source material for the Blake River assemblage. Both partial melting of tholeiitic to komatiitic material containing approximately 10 times chondritic REE abundances (Fowler and Jensen 1989) and very low degrees of partial melting of primitive mantle peridotite (Smith 1980) have been proposed.

Smith (1980) and Fowler and Jensen (1989) cited contrasting REE profiles between rhyolites of the Kinojevis North and South (flat profiles) and Blake River (LREE-enriched profiles) assemblages, with no patterns intermediate between the two, as evidence for a geochemical discontinuity between the Blake River and Kinojevis North assemblages. Neither Smith (1980) nor Fowler and Jensen (1989) demonstrated a continuum of magma source(es) and/or process(es) across the Blake River–Kinojevis North assemblage boundary, despite these assemblages either being coeval or having a maximum age difference of 4 Ma. The above observations suggest that the contacts between the Blake River and Kinojevis North and Kinojevis South assemblages are now fundamental structural and petrogenetic discontinuities.

Within Ontario, minor base metal mineralization occurs within the Blake River assemblage, in Ben Nevis Township, where the Interprovincial sulphide deposit (Canagau Mines Ltd.) contains galena, sphalerite, chalcopyrite, pyrite, pyrrhotite, gold and silver (Jensen 1975a). This mineralization occurs in veins and replacement zones within shear zones, which may represent a late structural localization, or may represent deformed early vein systems.

In Quebec, volcanic-associated massive base metal sulphide deposits of the Noranda area occur within a unit of chemically distinct felsic and andesitic metavolcanic rocks (Lesher et al. 1986). Metavolcanic rocks having this chemistry do not occur in the Blake River assemblage in Ontario (Lesher et al. 1986).

Gauthier Assemblage

The Gauthier assemblage (GA; *see* Figure 11.7) consists of ~2700 million-year-old (F. Corfu, Jack Satterly Geochronology Lab, Royal Ontario Museum, personal communication, 1990) felsic fragmental metavolcanic rock that crops out on the northern flank of the Timiskaming assemblage, within the core of the Spectacle Lakes Anticline (Thomson and Griffis 1944). The metavolcanic rocks consist mainly of coarse fragmental units (Thomson and Griffis 1944). The "agglomerate" members contain angular to subangular fragments ranging up to 2 m in length (Thomson and Griffis 1944). The Spectacle Lakes Anticline (Thomson 1943; Thomson and Griffis 1944) is doubly plunging and upright (Hamilton 1986; Hodgson and Hamilton 1989). The stratigraphic and/or structural relationships of the Gauthier assemblage, adjacent to the Timiskaming and Kinojevis South assemblages, are not known. The Gauthier assemblage is characterized by an aeromagnetically low signature.

Timiskaming and Hearst Assemblages

The belt of clastic metasedimentary rocks and associated alkalic metavolcanic rocks, centred on Kirkland Lake and extending east of Larder Lake (*see* Figure 11.7), is commonly referred to as the "Timiskaming" (herein called the Timiskaming assemblage). This is one of the best studied assemblages of the southern Abitibi greenstone belt. The Timiskaming assemblage (TI; *see* Figure 11.7) is important for several reasons: 1) it hosts some of the largest Archean lode gold deposits in the world; 2) it is the youngest Archean supracrustal unit of the Abitibi greenstone belt; and 3) it occupies a unique position in the tectonic framework of the southern Abitibi greenstone belt in that it postdates one regional deformation and predates the second.

An outstanding problem in the definition and interpretation of the Timiskaming assemblage is its relationship to turbiditic metasedimentary rocks south of the Larder–Cadillac Lake shear zone (herein called the Hearst assemblage).

Some workers have correlated the Timiskaming and Hearst assemblages (e.g., Hyde 1980; Hewitt 1963; Thomson 1946) and others have assigned them to different groups. Jensen (1978b) placed the Hearst assemblage (HE) metasedimentary rocks with mafic-ultramafic metavolcanic rocks of his Larder Lake group; however, the mapping of Hewitt (1963) and Thomson (1946) indicates that the Hearst assemblage is at least structurally disconformable with the Larder Lake assemblage and probably is unconformable.

The subdivision of metasedimentary rocks used here involves 2 distinct assemblages separated on the basis of facies association (e.g., Corfu et al., in press). Metasedimentary rocks of the resedimented facies are assigned to the Hearst assemblage. Metasedimentary rocks of the alluvial-fluvial facies are assigned to the Timiskaming assemblage.

The metasedimentary rocks of the Timiskaming assemblage are dominated by alluvial-fluvial facies association, consisting mainly of conglomerate (Figure 11.20) containing distinctive red chert (jasper) clasts and cross-bedded sandstone (Figure 11.21). These metasedimentary rocks are associated with distinctive alkalic ~2686 to 2677 million-year-old (Corfu et al., in press) metavolcanic rocks (Cooke and Moorhouse 1969; Hyde 1980). There are several outliers of metasedimentary rock north of the main belt of the Timiskaming assemblage (Thomson 1946). These outliers contain conglomerates bearing red chert clasts, trough cross-bedded sandstones and, in one case, alkalic metavolcanic rocks. These observations suggest that these outliers belong to the Timiskaming assemblage.

The Hearst assemblage is dominated by resedimented facies associations (Figure 11.22), mainly turbidites and, graded and/or matrix-supported conglomerate (Hyde 1980). These metasedimentary rocks are at least spatially associated with ~2705 million-year-old metavolcanic rocks of the Larder Lake assemblage. Metasedimentary rocks in the Hearst assemblage do not appear to be associated with alkalic metavolcanic rocks.

Although the Hearst and Timiskaming assemblages are dominated by resedimented facies and alluvial-fluvial facies, respectively, some similar units can be found in each assemblage. Transitional facies between the 2 main facies associations have not been observed (Hyde 1980) and where the 2 facies are adjacent, fault contacts are reported (Downes 1981; Hamilton 1986).

Conglomerates associated with the alluvial-fluvial facies of the Timiskaming assemblage are most commonly clast supported and have distinctive alkalic metavolcanic and intrusion clasts, red chert clasts, and "green carbonate" clasts (Bass 1961). The latter clast type may be derived from metavolcanic rocks altered prior to deposition of the Timiskaming assemblage. Conglomerates of the Hearst assemblage are most commonly matrix supported and do not commonly contain these clasts. Conglomerate clasts in the Hearst assemblage include green mica-bearing (fuchsite?) komatiite, rhyolite, chert, iron formation and carbonate fragments (Jensen 1985b).

Figure 11.20. Highly strained Timiskaming assemblage conglomerate, Gauthier Township. Banded nature of rock is imparted from highly strained clasts.

Figure 11.21. Interbedded Timiskaming assemblage clast-supported conglomerate and sandstone, near Chaput–Hughes on Highway 66.

The Timiskaming assemblage lies unconformably upon the previously deformed ~2700 million-year-old Kinojevis South assemblage (Thomson 1950; Hewitt 1963; Jackson 1988; *see* Figure 11.18). Unconformable contacts between the Hearst and previously deformed Larder Lake assemblages have also been described (Thomson 1946; Hewitt 1963). Thus, both the Hearst and Timiskaming assemblages represent late tectonic basins. They are not, however, necessarily coeval nor part of the same basin.

Figure 11.22. Matrix-supported conglomerate of the Hearst assemblage, southeastern Hearst Township.

Within the Timiskaming assemblage, there is a distinctive suite of alkalic metavolcanic rocks (Cooke and Moorhouse 1969). The alkalic rocks are commonly amygdaloidal and include: pseudoleucite-bearing flows, and trachyte flows, tuffs and agglomerates (Cooke and Moorhouse 1969; Thomson 1943). Calc-alkalic metavolcanic rocks are reported to occur near the base of the unit (Cooke and Moorhouse 1969). Phenocryst phases in the alkalic suite include pseudoleucite, olivine, augite to aegerine augite, alkali-feldspar, hornblende and biotite. Not all of these phenocryst types are present in all rocks.

In addition to high alkali content (Cooke and Moorhouse 1969; Basu et al. 1984; Ujike 1985), the alkalic rocks are characterized by extremely enriched LREE (Capdevila et al. 1982; Ujike 1985) and have positive $\varepsilon_{Nd}^{(t)}$ values (Basu et al. 1984) that tend to be lower than those of the pre–2700 Ma metavolcanic rocks of the southern Abitibi greenstone belt (Barrie and Shirey 1989).

The Timiskaming assemblage is a moderately to steeply south-dipping and south-facing assemblage that is cut by numerous faults and shear zones (Thomson 1946, 1950; Thomson et al. 1948; Hamilton 1986; Toogood and Hodgson 1986). The average dip of units within the Timiskaming assemblage is approximately 60°, but 20° to 45° dips are not uncommon (Thomson 1950). The approximately east-trending Larder–Cadillac shear zone is the most prominent regional structure, and in the eastern part of the Abitibi greenstone belt, it, in part, marks the boundary between the Timiskaming, Larder Lake and Hearst assemblages. Other prominent faults and/or shear zones in this region tend to be east-northeast to northeast or north striking, the latter being the latest fault set in the area. The east-northeast-striking faults include the well known Kirkland Lake fault ("Kirkland Lake main break").

The Kirkland Lake fault is a steeply south-dipping structure with a south-side-up sense of displacement that is well constrained (by piercing point analysis) to ~1500 feet at its western end and 350 feet at its eastern end (Thomson 1950; Thomson et al. 1950). The Kirkland Lake fault does not have a significant component of horizontal slip. Other faults subparallel to the Kirkland Lake fault do not significantly offset the Timiskaming–Kinojevis South assemblage contact, and thus, are also assumed to be predominantly dip-slip structures.

The Larder–Cadillac shear zone (also called the Larder Lake Break) is a steeply to moderately south-dipping structure that has ~2685 to 2675 million-year-old metavolcanic and metasedimentary rocks in the footwall and ~2705 million-year-old metavolcanic and metasedimentary rocks of the Larder Lake assemblage in the hanging wall. The fault is characterized by the common presence of talc-chlorite schist, intense carbonate alteration, quartz-veining and green mica. In some mine sections (e.g., Thomson 1943, 1948), the dip of the fault appears to shallow with increasing depth. Seismic reflection data (Green et al. 1990; Jackson, Sutcliffe et al. 1990) also suggest that this fault is south dipping (e.g., Jackson and Sutcliffe 1989). The Larder–Cadillac shear zone, therefore, appears to be a south-dipping fault with a significant thrust or reverse component of displacement (Thomson 1948; Hodgson and Hamilton 1989).

The Larder–Cadillac shear zone may be over 200 km long, extending from Kirkland Lake to Malartic and Cadillac in Quebec, and may be correlative with a similar structure near Matachewan (Powell et al. 1989). If a correlation between major linear belts of Timiskaming-like metasedimentary rocks and structures such as the Larder–Cadillac shear zone can be made, then the Larder–Cadillac shear zone, or similar structures, may extend through the Shining Tree and Swayze areas to the Ivanhoe Lake cataclastic zone, making the total strike length of the Larder–Cadillac shear zone, and/or related structures, over 350 km.

The most significant mineralization within the Timiskaming assemblage is gold, which is spatially related to shear zones, quartz veins and carbonate altered rocks. All of these are, in turn, at least spatially, related to some of the latest structures in the southern Abitibi greenstone belt, the Kirkland Lake fault ("Kirkland Lake main break") and the Larder–Cadillac shear zone ("Kirkland Lake break") (e.g., Todd 1928; Thomson et al. 1950). The Kirkland Lake fault, from which over 715 tonnes of gold have been extracted, is the locus of gold mineralization in the Lake Shore, Macassa, Teck–Hughes and Wright–Hargreaves mines. Spatially associated with the Larder–Cadillac shear zone are several major past-producing gold mines that collectively produced over 338 tonnes of gold (Kerr Addison Mine accounted for over 311 tonnes of gold). Reviews of gold mineralization in the Timiskaming assemblage include those of: Thomson et al. (1950), Charlewood (1964), Hodgson (1983a), Hodgson and Hamilton (1989), Downes (1981) and Smith et al. (1990).

Despite a general southward facing of units within the Timiskaming assemblage, the assemblage is structurally complex. The geochronological data of Corfu et al. (in press) indicate that the stratigraphy of the Timiskaming assemblage does not consistently young southwards and that faulting must be invoked to explain the age reversals. In addition, the abrupt termination of some alkalic metavolcanic members and the presence of a turbidite succession between alluvial-fluvial metasedimentary units (Dobie Basin of Toogood and Hodgson 1985) with no transitional facies between the two (Hyde 1980) indicates internal structural complexity of the Timiskaming assemblage.

The structure of the Hearst assemblage is discussed in the section entitled Larder Lake Assemblage.

Whitney Assemblage

The Whitney assemblage (WN; *see* Figure 11.7) consists largely of "...siltstones and minor feldspathic wackes in the basal part, and intermediate to mafic tuffs interlayered with minor siltstone in the upper part" (Pyke 1982). The assemblage lies between the Eldorado and Bowman assemblages, but the relationship of the Whitney assemblage to these assemblages is uncertain. The assemblage is truncated to the north by the Porcupine–Destor deformation zone, where it is juxtaposed against ultramafic metavolcanic rocks (Piroshco and Kettles 1988).

The assemblage is deformed about a shallowly plunging, northeast-striking anticline (Pyke 1982; Piroshco and Kettles 1988). Bedding orientations within the assemblage are commonly shallow to moderately dipping (Leahy 1971).

Contact relationships with adjacent assemblages, except to the north where the assemblage is truncated by the Porcupine–Destor deformation zone, are not known. The age of the assemblage is also unknown. Pyke (1982) and Leahy (1971) considered these metasedimentary rocks to be coeval with the "older" metavolcanic rocks (e.g., pre–2700 Ma).

Midlothian Assemblage

The Midlothian assemblage (MT; *see* Figure 11.7), consisting of interbedded turbidite, arkose, conglomerate, locally cross-bedded sandstone and lesser argillite, is situated at the eastern limit of the Halliday assemblage, west of Matachewan. Clasts within the conglomerate members include: dacite, rhyolite, quartz-feldspar porphyry, "greenstone", and white vein quartz. The assemblage crops out in 2 main areas: 1) in Halliday and Midlothian townships, the metasedimentary assemblage apparently lies disconformably on felsic metavolcanic rocks of the Halliday assemblage (Bright 1984); and 2) in Montrose Township, the assemblage lies between the Halliday assemblage and ultramafic to mafic metavolcanic rocks of the Bowman assemblage.

The disconformable relationship, presence of porphyry clasts and the cross-bedded nature of some sandstones suggests that the Midlothian assemblage is a lateral equivalent of the Timiskaming assemblage of the Kirkland–Larder Lakes area. If this interpretation is correct, then the Midlothian assemblage is likely ~2685 to 2675 Ma.

Bedding within the Midlothian assemblage is generally steeply dipping and faces to the north-northeast to north (Bright 1970, Map 2187). This contrasts with the general southward younging of the Timiskaming assemblage. Northeast-striking shear zones cut the boundary of the Halliday and Midlothian assemblages; however, significant displacement, in map view, is not apparent.

The Midlothian assemblage is unconformably overlain to the east by the Paleoproterozoic Huronian Supergroup.

ASSEMBLAGES IN THE ROUND LAKE AREA

Catharine–Pacaud Assemblage

The Catharine–Pacaud assemblage (CP), situated between the Round Lake batholith and the Skead assemblage (*see* Figure 11.7), consists of: the Pacaud structural complex, the Wawbewawa "formation", regionally extensive pyroxene-spinifex komatiite units, distinctive plagioclase-glomerophyric basalt flows, and the Catharine "formation" (Ridler 1969; Jensen 1985b; Jackson and Harrap 1989).

The Pacaud structural complex consists of compositionally banded mafic to intermediate rocks, foliated to massive mafic flows and/or sills, moderately to highly strained pillowed basalt, felsic (e.g., Jensen 1985b) to siliceous banded rocks, and a persistent lens of sulphide-facies iron formation adjacent to the Round Lake batholith (Lawton 1959; Jackson and Harrap 1989). Some units appear to be highly strained equivalents of the Wawbewawa formation (Jackson and Harrap 1989). The highly strained nature of this assemblage precludes correlation of its units. Some well-banded, highly strained rocks of the complex have been dated at ~2747 Ma (Mortensen, in prep.). This is one of the oldest ages in the southern Abitibi greenstone belt.

The Wawbewawa formation consists of non-amygdaloidal, magnesium-rich pillowed basalt, polysutured ultramafic rocks and massive mafic flows and/or gabbroic sills. At the top of the Wawbewawa formation is a regionally extensive, highly differentiated pyroxene-spinifex komatiite unit known as the Boston Creek komatiite (Stone et al. 1987; Jackson and Harrap 1989).

In Catharine Township, between the Boston Creek komatiite and the base of the Catharine formation, there is a succession of grey to green, massive to pillowed, locally amygdaloidal, basalt units and associated gabbro. This succession includes distinctive flows that are plagioclase-phenocrystic to glomerophyric (Jackson and Harrap 1989). Within this succession is another regionally extensive komatiite unit.

The Catharine formation is a thick unit of iron-rich, generally amygdaloidal, pillowed basalt (Figures 11.23 and 11.24). The exposed contact between the top of the Catharine–Pacaud assemblage and the overlying ~2700 million-year-old Skead assemblage displays both depositional and deformation features and is interpreted as a sheared primary contact (Jackson and Harrap 1989).

Southeast of the Round Lake batholith, Johns (1986) recognized a similar unit of metavolcanic rocks, with magnesium-rich tholeiitic basalt adjacent to the Round Lake batholith and iron-rich basalt between the magnesium-rich basalt and the Skead assemblage.

The units of the Catharine–Pacaud assemblage consistently face away from the Round Lake batholith (Lawton 1959; Grant 1963; Johns 1986; Jackson and Harrap 1989). Although the assemblage appears to be a simple homocline, Jackson and Harrap (1989) described evidence for a regional, near-layer-parallel fault that locally has a thrust-ramp geometry, although the fault itself may not be a thrust. This fault appears to curve around the batholith and may correlate with shearing at the base of the Catharine formation near Trestle Hill, west of Englehart (*see* map *in* Johns 1986).

Metavolcanic rocks adjacent to the Round Lake batholith do not display evidence of significantly greater metamorphic grade than the regional greenschist facies assemblages found farther away from the batholith. Locally, amphibolite facies conditions are recorded (Jolly 1980; Johns 1986); however, they are generally restricted to less than 500 to 1000 m from the batholith margin, which suggests a near-vertical contact or "cold" emplacement of the batholith.

Penetrative fabric within the Catharine–Pacaud assemblage is most pronounced adjacent to the Round Lake batholith and is locally developed near faults. Penetrative fabrics are difficult to discern elsewhere in the assemblage.

Goodwin (1979) and Johns (1986) demonstrated that metavolcanic rocks of the Catharine–Pacaud assemblage belong to a tholeiitic suite. Capdevila et al. (1982) reported slightly LREE-depleted to slightly LREE-enriched, fractionated (1 to 20 times chondrite) REE patterns for the Catharine–Pacaud assemblage. These REE patterns suggest that basalt units of the Catharine–Pacaud assemblage were derived by partial melting of a variously depleted mantle source. No evolved crustal contribution is suggested by the geochemical data (Capdevila et al. 1982).

Mineralization within the Catharine–Pacaud assemblage includes: 1) gold associated with shears, quartz veins, carbonate alteration (Burrows and Hopkins 1922, 1925; Bell 1930; Ridler 1970) and the northwest-striking brittle to brittle-ductile Catharine fault (Jackson and Harrap 1989); and 2) copper-sulphide mineralization associated with the sulphide layer that borders the Round Lake batholith (Bell 1930).

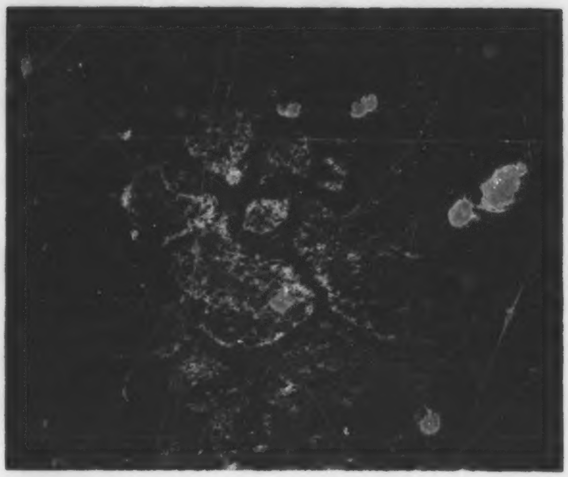


Figure 11.23. Amygdaloidal, pillowed, tholeiitic basalt of the Catharine formation from the Catharine-Pacaud assemblage, Catharine Township. The pocket knife in upper left-hand corner indicates scale.

Figure 11.24. Variolitic basalt of the Catharine formation near the Catharine shear zone, Catharine Township. The pen in the upper left-hand corner indicates scale.

Skead Assemblage

The ~2701 million-year-old (Corfu et al. 1989) Skead assemblage (SK; *see* Figure 11.7) is composed predominantly of calc-alkalic, felsic to intermediate metavolcanic rocks (e.g., Goodwin 1979). The assemblage is bounded to the: 1) southwest by the Catharine–Pacaud assemblage; 2) northeast by the McElroy assemblage; 3) west by the Boston assemblage; and 4) east by the Paleoproterozoic Huronian Supergroup. On a regional scale, the assemblage wraps around the eastern and southern margin of the Round Lake batholith (Jensen and Langford 1985; Johns 1986); units face away from the batholith (e.g., Hewitt 1951; Johns 1986). The Skead assemblage metavolcanic rocks can be subdivided into a unit of fragmental rock composed of dacitic to rhyolitic tuff, breccia and polylithic agglomerate (Figure 11.25), and a second unit of predominantly dacitic, massive to fragmental, metavolcanic rocks (Hewitt 1951).

Figure 11.25. Polylithic, intermediate, fragmental metavolcanic rock of the Skead assemblage, Skead Township. Note the angular nature of fragments, the variation in fragment size and the fine-grained matrix.

East of St. Anthony Lake, there is a high concentration of subvolcanic porphyritic intrusions; bedding is poorly developed and the fragments range from cobble to boulder size (Hewitt 1951; Jensen 1985b). To the northeast, however, the fragment size decreases and bedding becomes well developed. These relationships suggest that the St. Anthony Lake area represents a relatively proximal volcanic environment; the northwestern part of the Skead assemblage represents a more distal environment (Jensen 1985b).

The Skead assemblage is characterized by fractionated, LREE-enriched patterns with no europium anomaly (Capdevila et al. 1982). Based on the REE patterns and chromium, nickel, yttrium and scandium contents, Capdevila et al. (1982) concluded that metavolcanic rocks of the Skead assemblage were derived by moderate to low degrees of melting of undepleted mantle peridotite.

The basal or southern contact of the Skead assemblage with the Catharine–Pacaud assemblage was interpreted by Jackson and Harrap (1989) as a sheared primary contact. If the base of the Skead assemblage is similar in age to the ~2700 million-year-old unit dated by Corfu et al. (1989), then there is an approximate 20 million-year gap in volcanic activity between the Skead and Catharine–Pacaud assemblages.

Most metavolcanic rocks of the Skead assemblage display little or no penetrative fabric. However, schistosity is recorded south of, and subparallel to, the Lincoln–Nipissing shear zone. Apparent corridors of schistosity (*see* Hewitt 1951, Map 1949–3) include: 1) a northeast-trending zone along strike from northeast-striking faults located near the northeast arm of Larder Lake; and 2) an east-trending zone of schistosity located northwest and north of St. Anthony Lake. Northeast- and east-striking faults do not display significant displacement in map view.

The Skead assemblage is not richly mineralized; however, nearly 0.311 metric tonnes of gold were produced from the Cathroy–Larder Mine, which is located near the contact between the Skead and Larder Lake assemblages (Abraham 1951; Meyer et al. 1989). The mine is situated in highly fractured, sheared, silicified and carbonatized felsic fragmental rocks that are cut by numerous, thin, sulphide seams.

McElroy Assemblage

The McElroy assemblage (ME; *see* Figure 11.7) is a northwest-trending homoclinal succession of massive to pillowed mafic flows and minor felsic fragmental units (Hewitt 1951; Abraham 1951; Jackson, Kimmerly et al. 1990). Many of the mafic rocks are medium to coarse grained and are referred to by intrusive names rather than extrusive names (e.g., gabbro versus basalt); however, the intimate relationship of these rocks with pillowed mafic metavolcanic rocks and felsic fragmental units may indicate that they are of effusive origin.

Medium- to coarse-grained mafic rocks of the McElroy assemblage include: 1) hornblende gabbro, commonly with well-developed dendritic patterns of polycrystalline hornblende that locally resembles pyroxene-spinifex texture; 2) leucogabbro; 3) gabbro; and 4) plagioclase-glomerophyric gabbro. Fine- to medium-grained mafic to ultramafic rocks include pillowed and locally spinifex-textured and polysutured units.

The nature of the contact between the Skead assemblage and McElroy assemblage is not known. The boundary between the McElroy and Larder Lake assemblages is marked by: 1) the Lincoln–Nipisssing shear zone; 2) the Manor shear zone; and 3) the Lincoln–Nipissing peridotite. Although similar rock types exist in both assemblages, the proportion of units differs dramatically. The McElroy assemblage is dominated by massive mafic rocks, whereas the Larder Lake assemblage is dominated by pillowed mafic volcanic rocks.

Larder Lake Assemblage

The Larder Lake assemblage (LL; *see* Figure 11.7), as here defined, consists predominantly of metavolcanic rocks and only minor metasedimentary rocks. This assemblage is bounded to the: 1) north by the Timiskaming assemblage along the Larder–Cadillac shear zone; 2) south by the McElroy assemblage; 3) west by the Boston assemblage along the Boston fault; and 4) east by the Paleoproterozoic Huronian Supergroup. The Larder Lake assemblage is also structurally overlain by the metasedimentary Hearst assemblage (*see* Hearst Assemblage).

Metavolcanic rocks of the Larder Lake assemblage consist predominantly of tholeiitic basalt, komatiitic basalt and ultramafic komatiite (Jensen 1985a, 1985b; Figure 11.26). North of the Lincoln–Nipissing shear zone, many of the basalt units are pillowed and felsic metavolcanic rocks are uncommon, except just northeast of the junction of the Lincoln–Nipissing peridotite and the Huronian Supergroup. Some turbiditic metasedimentary rocks are interbedded with komatiite (Jensen 1985a, 1985b), but the maps of Hewitt (1951) and Thomson (1949) indicate that most of the metasedimentary rocks either rest unconformably upon the metavolcanic rocks or are in profound structural discordance with the metavolcanic rocks.

The Larder Lake assemblage, as defined here, is less extensive than the Larder Lake group of Jensen (1985b) and

Figure 11.26. Polysutured ultramafic flow of the Larder Lake assemblage, McElroy Township.

does not include: 1) most of the metasedimentary rocks between the Lincoln–Nipissing shear zone and the Larder–Cadillac shear zone, which appear to unconformably overlie metavolcanic rocks of the Larder Lake assemblage (Thomson 1949; Hewitt 1951, 1963) and thus are presumed to belong to the Hearst assemblage; and 2) the iron formation bearing Boston assemblage, which is in fault contact with the Larder Lake assemblage. Felsic metavolcanic rocks, believed to be part of the Larder Lake assemblage, have been dated at ~2705 Ma (Corfu et al. 1989).

The Larder Lake assemblage has been subject to both pre– and post–Timiskaming folding. The post–Timiskaming folds are predominantly west to northwest striking, although some folds are northeast trending and may be related to a fault system that extends southwesterly from the northeast arm of Larder Lake (e.g., Hamilton 1986).

Numerous faults transect the Larder Lake assemblage; these include northwest-striking faults that are cut by both north- to west-northwest-striking faults and east-northeast- to northeast-striking faults.

The Manor shear zone and the more southerly Lincoln–Nipissing shear zone juxtapose Skead (~2701 Ma) and Larder Lake (~2705 Ma) assemblages (e.g., Corfu et al. 1989). The facing and age relationships of these assemblages suggest a component of north-side-up displacement; however, the kinematics are not understood and lateral juxtaposition of assemblages may equally well explain the data. Alternatively, a south-side-up displacement could be interpreted because the Hearst assemblage appears to be younger than the Skead, McElroy and Larder Lake assemblages in that some conglomerate clasts in the Hearst assemblage are comparable to units found in the Skead, McElroy and Larder Lake assemblages to the south of the fault.

The Lincoln–Nipissing shear zone is truncated by the Boston fault, which in turn bends around the Lebel Stock. Therefore, the Boston fault is pre- to syn-emplacement of the Lebel Stock; consequently, the Lincoln–Nipissing shear zone is of similar age. Aeromagnetic maps suggest that the Lincoln–Nipissing shear zone terminates the southern margin of the McElroy Stock. These observations suggest that the Lincoln–Nipissing shear zone was active around the time of emplacement of the alkalic plutons.

Numerous intrusions cut the Larder Lake assemblage; these include many small quartz ± porphyry bodies and larger syenitic bodies, such as the McElroy Stock.

Boston Assemblage

The Boston assemblage (BS; *see* Figure 11.7) consists of metavolcanic and metasedimentary rocks which include komatiite, tholeiitic basalt, serpentinized ultramafic rocks, oxide and sulphide facies iron formation, chert, conglomerate, sandstones and felsic metavolcanic rocks (Abraham 1951; Lawton 1959; Lovell 1972; Jensen and Langford 1985). The Boston assemblage (*see* Figure 11.7) abuts against the Timiskaming assemblage along the Larder–Cadillac shear zone (*see* Timiskaming Assemblage). The southern margin of the assemblage is delineated by the Boston fault and the Round Lake batholith. The aeromagnetic pattern of this assemblage is complex because of lithologic and structural complexity. As here defined, the Boston assemblage corresponds to that part of the Larder Lake group described by Jensen (1985a, 1985b) that lies west of the Boston fault. The Boston assemblage is best preserved between the eastern margin of the Lebel Stock and the Boston fault. The assemblage also extends around the southern and western margin of the Lebel Stock and is here correlated with rocks north and west of the Otto Stock towards Matachewan that include komatiite (Jensen 1978b), iron formation, pillowed basalt and felsic metavolcanic rocks.

The association of iron formation and felsic and ultramafic rocks in the Boston assemblage is similar to the Eldorado and Bartlett assemblages, but dissimilar to the Skead, Blake River and Halliday assemblages. Furthermore, the aeromagnetic pattern for the Boston assemblage is similar to the Eldorado and Bartlett assemblages, but contrasts with the flat pattern of the Skead, Blake River and Halliday assemblages. At the time of writing, it was not possible to constrain the age of this assemblage, but it may be similar to that of the 2727 million-year-old Bartlett assemblage.

The Adams Iron Mine (Dubuc 1966), located east of the Lebel Stock, has produced over 22 675 tonnes of iron ore (Meyer et al. 1989). The iron range consists of interbedded chert- and magnetite-rich layers (Dubuc 1966; Figure 11.27). Other mineralization within the Boston assemblage includes the Dane copper mine and some minor gold mineralization near the Larder–Cadillac shear zone, which marks the boundary between the Boston and Timiskaming assemblages (Thomson 1950).

Within the Boston assemblage, rock units and well-developed foliations are mainly steeply dipping and subparallel to the syenite stocks and the Larder–Cadillac shear zone. Numerous folds are outlined on township maps (Lawton 1959), but their plunges are not generally known. East of the Lebel Stock, folds of the iron formation plunge approximately 60° to the southeast (Dubuc 1966).

The Boston fault juxtaposes the Boston assemblage against the Larder Lake, Skead and Catharine–Pacaud assemblages. Aeromagnetic data suggest that the Boston

Figure 11.27. Banded iron formation of the Boston assemblage, Boston Township. Light-coloured bands are silica rich and dark-coloured bands are magnetite rich.

fault and the northwestern extension of the fault at the base of the Catharine formation (*see* Catharine–Pacaud Assemblage) merge at the northeastern corner of the Otto Stock. This fault has not been recognized north and west of the Otto Stock; however, it may project into this region as a subtle, strata-parallel fault.

The Boston fault formed after 2700 Ma because it cuts the ~2701 million-year-old Skead assemblage. It may predate, or have formed at, ~2680 Ma, since it is concentric about the southeastern margin of the Lebel Stock (likely coeval with the Otto Stock, ~2680 Ma, Corfu et al. 1989). The Lebel Stock predates the Larder–Cadillac shear zone, so the Boston fault must also predate the Larder–Cadillac shear zone.

ASSEMBLAGES IN THE SHINING TREE AREA

The Shining Tree area (*see* Figures 11.6 and 11.7) is bounded to the: 1) northwest by the Kenogamissi Batholith; 2) southwest by the Ramsey–Algoma granitoid complex; and 3) east by unconformably overlying Paleoproterozoic rocks of the Huronian Supergroup. Carter (1983, 1987) mapped the Shining Tree area prior to developments in lithologic mapping in the main part of the Abitibi greenstone belt that resulted in recognition of the main lithostratigraphic suites. The Exploration Division of Kidd Creek Mines Limited remapped the Shining Tree area and subdivided the area into "groups" similar to those of the main Abitibi greenstone belt (MERQ–OGS 1983). The relationship of these groups to other units of the Abitibi greenstone belt is uncertain. The following description of the area is slightly modified after the description provided by MERQ–OGS (1983).

Natal Assemblage

The Natal assemblage (NT; *see* Figure 11.7) consists of "turbiditic wackes" (MERQ–OGS 1983, Natal group) with flows of komatiitic to tholeiitic affinity (MERQ–OGS 1983, Bigfour group) in the central part of the Shining Tree area and with calc-alkalic metavolcanic rocks (MERQ–OGS 1983, Upper Deloro group) in the southeastern part of the Shining Tree area.

Cabot–Kelvin, Shining Tree and Sinclair Assemblages

The remaining units of the Shining Tree area are described by MERQ–OGS (1983) as follows: "An unconformity separates this sequence [Natal and Ridout assemblages] from the Cabot–Kelvin [CK] group [assemblage] of calc-alkalic metavolcanics underlain by Shining Tree [ST] group [assemblage] tholeiites, Sinclair [SC] group [assemblage] komatiites...".

Ridout Assemblage

The youngest assemblage in the Shining Tree area is the Ridout assemblage (RO; *see* Figure 11.7) which contains jasper-pebble conglomerate, "turbiditic wackes" and alkalic flows. The Ridout assemblage lies unconformably on top of adjacent assemblages, separating a south-facing, northeast-trending, ultramafic to felsic metavolcanic sequence to the north from a northeast-facing, northwest-trending, ultramafic to felsic metavolcanic sequence to the south. The Ridout assemblage extends west across the Swayze area for approximately 100 km.

In the Swayze area, the Ridout assemblage consists of a narrow band (less than 2 km) of steeply dipping turbidites, arkose and conglomerate, containing well-rounded pebbles and boulders of "granite", chert, vein quartz, mafic metavolcanic rock, porphyritic rhyolite and rare jasper fragments (Furse 1932; Rickaby 1935; Laird 1936a; Meen 1944; Siragusa 1987). Units of metavolcanic rock and iron formation are locally interlayered with the clastic metasedimentary rocks (Rickaby 1935; Meen 1944; Donovan 1968). The presence of interlayered iron formation may indicate that some units, which historically have been assigned to the Ridout assemblage, may represent intermediate to felsic, fragmental, metavolcanic rocks. Alternatively, the contact with the adjacent metavolcanic rocks may be more structurally complex.

In the western part of the Swayze area, bedding is generally poorly developed in the conglomerates, cross-bedding is not observed and facing directions are not well constrained (Donovan 1968). In the central part of the Swayze area, turbidite and arkose units are well bedded and have an easterly to northeasterly strike. These units are locally cross-bedded (Meen 1944, p.14) and face to the north, but are overturned to the south (ca. 70°; Furse 1932). In Benton Township, a chert-jasper iron formation unit is apparently interlayered with clastic metasedimentary units, interpreted to belong to the Ridout assemblage (Laird 1936a). Unlike other iron formations in this part of the Swayze area, this iron formation unit is jasper rich (Laird 1936a).

The Ridout assemblage is widest (up to 4 km in map view) in the eastern part of the Swayze area (Opeepeesway Lake area). To the east of the Swayze area, it thins to less than 0.5 km in map width. Conglomerate, turbidite and arkose are the principal rock types; conglomerate pebbles consist of "granite", iron formation, mafic metavolcanic fragments, quartz and porphyry (Laird 1936a, 1936b; Moorhouse 1951). Cross-bedding in arkosic units is reported; units face north, but locally dip steeply to the south (Laird 1936a; Moorhouse 1951). Bedding trends are grossly parallel with stratigraphic trends in the adjacent metavolcanic assemblages; however, intense shearing in the metavolcanic units prohibits a more rigorous evaluation of bedding discordances. East-striking shearing is also localized within the metasedimentary assemblage, and locally on both north and south contacts with the adjacent metavolcanic rocks (Moorhouse 1951). Bodies of lamprophyre and quartz-feldspar and red feldspar (syenite) porphyry have intruded the metasedimentary rocks (Moorhouse 1951). The presence of albitic phenocrysts and attendant hematite alteration (Moorhouse 1951) suggests that some of the late porphyry bodies have alkalic affinities.

The Ridout assemblage was considered equivalent in age with the Keewatin metavolcanic rocks (Emmons and Thomson 1929). However, based on the presence of "granite" and jasper fragments, cross-bedding in the turbidites (Laird 1932, 1936a; Rickaby 1935), alkalic flows in the Shining Tree area and the inferred unconformable relationship between the metasedimentary and adjacent metavolcanic rocks (Meen 1944), it is likely that the Ridout assemblage is a temporal and tectonic equivalent of the Timiskaming assemblage in the Kirkland Lake area (e.g., Meen 1944; Moorhouse 1951). Locally, intense shearing is localized along the contacts between the metasedimentary and metavolcanic rocks (Rickaby 1935), which may account for the difficulty in recognizing the original stratigraphic relationship with the metavolcanic rocks (e.g., Laird 1936a; Moorhouse 1951, p.10; Siragusa 1987).

The past-producing Tyranite gold mine in the Shining Tree assemblage, located in northern Tyrrell Township (Carter 1987), is the only known significant mineralization within the Shining Tree area. The gold is associated with carbonatized shear zones. Other significant gold occurrences include the Jerome gold mine (Moorhouse 1951), which occurs along a southeasterly striking shear zone in the Opeepeesway Lake area (Siragusa 1987).

ASSEMBLAGES IN THE SWAYZE AREA

The Swayze area (*see* Figure 11.6) lies west of the Kenogamissi Batholith and east of the Kapuskasing Structural Zone. Units east of the Kenogamissi Batholith are connected to greenstones of the Swayze area by thin metavolcanic units which wrap around the north and south edges of the Kenogamissi Batholith. Numerous north-northwest-striking faults, having a sinistral displacement sense, cut across the area. These faults cut all metavolcanic and metasedimentary rocks, as well as the Kenogamissi Batholith. Three Proterozoic diabase dike swarms are present: 1) the north-striking Matachewan dike swarm; 2) the northwest-striking Sudbury dike swarm; and 3) the east- to northeast-striking Abitibi dike swarm (Fahrig and West 1986; *see also* Osmani, this volume).

Muskego–Reeves Assemblage

The Muskego–Reeves assemblage (MR) is a greenschist to amphibolite facies, east-striking, steeply dipping, lithologically heterogeneous assemblage (composite) that abuts a biotite trondhjemite gneiss and biotite-hornblende granodiorite (*see* Figure 11.7) to the north (Milne 1972). The western boundary is defined by the Kapuskasing Structural Zone and a massive granodiorite to granite intrusion. To the east, the assemblage thins considerably, but may pass north of the Kenogamissi Batholith (*see* Figure 11.7). About 60% of the assemblage consists of pillowed, amygdaloidal basalt (tholeiitic?) flows (Milne 1972). Tremolite-rich basalts (Milne 1972) may correspond to komatiitic basalts or magnesium-rich basalt. Amphibolitized basalts occur adjacent to the trondhjemitic and granodioritic intrusions (Milne 1972). Komatiitic flows and related ultramafic and gabbroic intrusions are associated with units of oxide facies iron formation, clastic metasedimentary rocks and felsic metavolcanic rocks in the central part of the assemblage (Milne 1972; J.C. Ireland, Porcupine Resident Geologist's office, personal communication, 1990). Komatiitic units have been described north of the Kenogamissi Batholith (Choudry 1989). Given the diverse rock types in this assemblage, subdivision into additional assemblages, including possible Timiskaming-like assemblages, may be possible following further mapping.

The felsic metavolcanic rocks consist of quartz- and feldspar-phyric flows, tuffs, lapilli tuffs and breccia (Prest 1951; Milne 1972). The presence of scour and graded bedding suggest that the felsic metavolcanic unit has been locally reworked (Milne 1972) and that the unit contains "tuffaceous wacke".

A steeply dipping unit of clastic metasedimentary rock, interpreted to be conformable with the associated volcanic deposits, consists of interlayered turbidite and conglomerate (Prest 1951; Milne 1972). Conglomerate pebbles are rounded to angular and include porphyritic "granite", aphanitic felsic fragments and diabase-textured basalt (Milne 1972, p.14). The unit is intruded by concordant bodies of feldspar porphyry (Milne 1972). The clast-types suggest that the sediments are Timiskaming-like, however, the nature of these metasedimentary rocks has not been confirmed.

Minor occurrences of chert-magnetite, carbonaceous sulphide (pyrite) and carbonate-bearing iron formation are associated with felsic metavolcanic rocks (Milne 1972). The Radio Hill iron formation (Milne 1972), which strikes easterly, dips moderately to steeply (50° to 90°) to the north and overlies a schistose felsic metavolcanic unit, consists of chert, magnetite, siderite, minnesotaite and pyritic carbonaceous layers (Milne 1972).

An east-trending corridor of shallow-dipping to subhorizontal foliation is localized in the central part of the assemblage, and corresponds to the western extension of the Porcupine–Destor deformation zone (Milne 1972). Other east-striking shear zones, some of which correspond to zones of back-to-back facing directions, are localized within the assemblage (Prest 1951; Milne 1972). Extensive replacement by hydrothermal carbonate has taken place within and adjacent to deformation zones (Milne 1972). Elsewhere, a well-developed, east-striking foliation has affected all rocks (Milne 1972).

Talc and asbestos have been extracted from bodies of serpentinized and carbonatized ultramafic (komatiitic) rock (Milne 1972). Significant gold occurrences (e.g., Joburke Mine) are associated with several east-striking shear zones (Milne 1972).

Hanrahan Assemblage

The Hanrahan assemblage (HN), a greenschist to amphibolite facies assemblage in the northeastern part of the Swayze area (*see* Figure 11.7), is intruded by the Hardiman Lake Pluton, which is a foliated, biotite-hornblende, trondhjemite-tonalitic phase of the Kenogamissi Batholith (Milne 1972; Breaks 1978). The assemblage consists of intermediate to felsic, feldspar-phyric, fragmental metavolcanic rocks, some of which carry pumiceous fragments (Milne 1972). The assemblage is cut by mafic and ultramafic intrusions (Milne 1972).

Oxide-facies iron formation (1 to 30 m thick), which consists of chert, magnetite and iron-rich silicate (amphibole or chlorite), and rarely pyritic carbonaceous material, caps the felsic metavolcanic unit (Milne 1972). Along the northwestern edge of the assemblage, the iron formation has a steep, northwesterly dip (Milne 1972). Hydrothermal sulphide minerals have been concentrated where the iron formation is folded (Milne 1972).

A well-developed, east-striking foliation obscures most primary features in the metavolcanic rocks. This foliation is axial planar to a 65°, northwest-plunging anticline, which folds the assemblage (Milne 1972). The contact with the Muskego–Reeves assemblage to the west is not exposed.

Horwood Assemblage

The arcuate Horwood assemblage (HW) extends from the Ivanhoe Lake area in the northwest to Raney Township in the southwest (*see* Figure 11.7). To the east, the assemblage has been intruded by the Hardiman Lake Pluton (Breaks 1978). This greenschist facies assemblage consists principally of massive, pillowed and amygdaloidal, iron-rich, tholeiitic basalt, with associated flow breccia (Breaks 1978; J.C. Ireland, Porcupine Resident Geologist's office, personal communication, 1990). Few units of iron formation or carbonaceous interflow metasedimentary rocks occur, based on the absence of airborne electromagnetic (AEM) conductors in the southern half of the assemblage (OGS 1982). Variolitic flows have been identified in the Horwood Lake area (Breaks 1978, p.10). Basalt flows in the Horwood Lake area face westerly, away from the Hardiman Lake Pluton (Breaks 1978). Locally, amphibolitized, laminated basalt, possessing a shallow dipping (less than 60°) foliation, occurs adjacent to the Hardiman Lake Pluton (Breaks 1978). Mafic and ultramafic intrusions and a few units of dacitic and rhyolitic flows, lapilli tuff, pyroclastic breccia and associated turbidite occur interlayered with the basalts (Breaks 1978).

Several shear zones occur in the northeastern part of the assemblage, near Horwood Lake. Several gold occurrences and numerous ultramafic and felsic intrusions occur near the northeast-striking Hardiman shear zone (Milne and Breaks 1972; Breaks 1978). Oblique, sinistral displacement along the zone is suggested by northeast-plunging (45°), S-shaped folds and deflected foliation trajectories. A second, northeast-trending shear zone, which may also have oblique sinistral displacement, is inferred to occur immediately north of the Hardiman shear zone. The north-trending Horwood–Hoodoo fault separates south-facing basalts to the east from west- to southwest-facing pillow basalts to the west (Breaks 1978).

To the north of the Hardiman shear zone, an elongate, massive, magnetite-bearing, biotite-hornblende quartz diorite and biotite-hornblende granodiorite pluton crops out (Horwood peninsula pluton; Breaks 1978, p.36). The Hoodoo Lake pluton, a potassium feldspar-megacrystic, biotite granodiorite, crops out northwest of Horwood Lake (Breaks 1978). The massive Kukatush stock, localized along the boundary between the Horwood Lake and Muskego–Reeves assemblages, consists of hornblende diorite and hornblende quartz monzonite (Milne 1972; Breaks 1978). Quartz-rich phases occur towards the outer margins of this pluton (Breaks 1978).

Numerous gold occurrences are spatially associated with the northeast-striking Hardiman shear zone and its possible southwesterly extension from the Horwood assemblage into the Raney–Newton assemblage.

Raney–Newton Assemblage

The Raney–Newton assemblage (RN) is a wedge-shaped assemblage (*see* Figure 11.7) that is thickest (10 km) in the east, against the Kenogamissi Batholith, and thins to the southwest. The assemblage consists of east-striking, south-facing, steeply dipping calc-alkalic andesitic, dacitic and rhyolitic flows, pyroclastic rocks and associated clastic metasedimentary rocks (Furse 1932; Rickaby 1935; Laird 1936a, 1936b; Moorhouse 1951; Donovan 1965, 1968; Thurston et al. 1977; J.C. Ireland, Porcupine Resident Geologist's office, personal communication, 1990). The clastic rocks and minor associated 2697 million-year-old (Cattell et al. 1984) felsic metavolcanic rocks along the southern, or top, part of the Raney–Newton assemblage, have historically been called the "Swayze series" (Furse 1932). In this southern part of the assemblage, felsic metavolcanic breccia is more abundant than tuff (Donovan 1965). Massive, porphyritic, felsic metavolcanic rock may represent either an effusive volcanic facies (Donovan 1965, 1968) or subvolcanic intrusions (Rickaby 1935). A continuous, east-striking, airborne electromagnetic (AEM) conductor occurs along the southern contact of the assemblage (OGS 1982).

Basalt in the northeastern part of the assemblage is greenschist facies (Breaks 1978), except adjacent to the Kenogamissi Batholith, where amphibolite facies was attained (Breaks 1978). A well-developed, steeply dipping penetrative foliation is present in all metavolcanic rocks.

West of Rollo Lake, the assemblage thins to less than 5 km in map view and is truncated by the Kapuskasing Structural Zone. In this area, the assemblage consists of andesitic and rhyolitic flows and pyroclastic rocks along the north edge, and interlayered felsic metavolcanic and clastic metasedimentary rocks of the "Swayze series" along the south edge (Rickaby 1935; Donovan 1968).

"SWAYZE SERIES"

The "Swayze series" forms the southern part of the Raney–Newton assemblage and consists of conglomerate, turbiditic wacke, arkose and chert which are locally interlayered with felsic metavolcanic rocks (Furse 1932; Rickaby 1935). In the western and central part of the Raney–Newton assemblage, the metasedimentary rocks face south, but are overturned and dip steeply (60° to 80°) to the north (Furse 1932, p.42). Conglomerate pebbles consist of felsic porphyry, chert and mafic metavolcanic rock, but significantly, no "granite" pebbles occur (Furse 1932, p.43). Based on the pebble types, Furse (1932) concluded that the metasedimentary rocks of the "Swayze series" were not of Timiskaming-type and were not correlative with the Ridout assemblage, located to the south.

The "Swayze series" is interlayered with metavolcanic rocks along the southern part of the Raney–Newton assemblage (Rickaby 1935). There, metavolcanic fragment types predominate in the metasedimentary rocks (Furse 1932; Rickaby 1935; Donovan 1965). These observations are consistent with a volcanic and clastic origin for the "Swayze series", which appears to be stratigraphically equivalent to the felsic metavolcanic rocks (Thurston et al. 1977) of the Raney–Newton assemblage.

The location of the northern Raney–Newton assemblage contact is approximate, based on the contrasting magnetic signatures between rocks of the Horwood assemblage to the north (OGS 1982). The inferred position of the contact with the Horwood assemblage corresponds approximately with the axial trace of an anticline reported by Thurston et al. (1977). However, since the assemblages are separated by the axial trace and are in a back-to-back facing relationship, the contact between the Horwood and Raney–Newton assemblages may alternatively be a fault.

In the eastern part of the assemblage, the southern boundary is defined by an east-trending AEM conductor and associated "Swayze series" metasedimentary rocks. The apparent wedge shape of the Raney–Newton assemblage, the possible age inversion across the contact (e.g., Cattell et al. 1984; *see* Halcrow–Swayze Assemblage) and the opposite facing directions across the contact (Donovan 1965, 1968) are consistent with a structurally complex southern contact. A northwest-striking shear zone occurs along this contact in the Crossley Lake area (e.g., Rickaby 1935, p.20).

Halcrow–Swayze Assemblage

The Halcrow–Swayze assemblage (HS) is an east-trending, greenschist to amphibolite facies assemblage (*see* Figure 11.7) which consists of komatiitic flows, tholeiitic basalt, intermediate to felsic and calc-alkalic metavolcanic rocks, interlayered with oxide-facies iron formation (Donovan 1965, 1968; Goodwin 1965; Thurston et al. 1977; J.C. Ireland, Porcupine Resident Geologist's office, personal communication, 1990). To the east, the assemblage splits into 2 bands which lie north and south of the Marion assemblage (*see* Marion Assemblage). Titanium oxide-rich (greater than 1.2 weight %), pillowed and flow breccia andesite, interlayered with numerous, thin, felsic pyroclastic units, occurs to the west of the Woman River iron formation (Goodwin 1965, p.12, 26). Amphibolitic tholeiitic basalt occurs along the south limb, adjacent to the southwest margin of the Kenogamissi Batholith (Moorhouse 1951; Siragusa 1987). These flows are overlain by intermediate and felsic, calc-alkalic tuffs and breccias located north of Opeepeesway Lake (Moorhouse 1951; Siragusa 1987). East- to southeast-striking shearing is intense along this south limb (Moorhouse 1951). Mineral and stretching lineations, related to this shear, have a shallow (30°), westerly plunge (Moorhouse 1951).

In the western part of the assemblage, tholeiitic basalt and komatiitic flows occur preferentially near the assemblage margins; intermediate to felsic metavolcanic rocks are concentrated in the core (J.C. Ireland, Porcupine Resident Geologist's office, personal communication, 1990). Based on few data, this unit faces north (Donovan 1968). An aureole of amphibolite facies contact metamorphism is developed at the western edge (Donovan 1968), where the assemblage is truncated by the Ivanhoe Lake cataclastic zone (Thurston et al. 1977).

Along the northern edge of the assemblage, komatiitic and tholeiitic flows crop out immediately south of the Raney–Newton assemblage. Tholeiitic basalts have flat REE patterns and relatively low MgO, Ni, CaO/TiO_2, and high Ti/V values (Cattell et al. 1984). The following types of komatiites are present (Cattell et al. 1984): 1) alumina-, LREE- and HREE-depleted, characterized by low Al_2O_3/TiO_2, high CaO/Al_2O_3; and 2) alumina- and LREE-depleted, characterized by chondritic incompatible element ratios and flat HREE. Three parental magmas may account for the observed chemical variations (Cattell et al. 1984; Cattell and Arndt 1987): 1) a komatiite magma depleted in both LREE and HREE; 2) LREE-enriched magma with sloping HREE; and 3) LREE-depleted magma with flat HREE.

The komatiites along the northern contact of the assemblage are continuous along strike for about 40 km and are characterized by a high magnetic signature (OGS 1982). This unit has been tightly folded about an east-northeast-striking syncline, which has a steeply dipping axial plane (Donovan 1965, 1968; Goodwin 1965). The syncline appears to close both to the west and east and may, therefore, be shallowly plunging (OGS 1982).

Two units of felsic pyroclastic rock occur associated with the northern komatiite unit. Both these felsic metavolcanic rocks consist of massive and pyroclastic units, similar to those described in the Raney–Newton assemblage (Donovan 1965, 1968; Goodwin 1965). The thickest (3 km) felsic unit occurs in the Brett Lake area in Swayze

Township, on the north limb of the syncline (Donovan 1965). It lies south of the northern komatiite unit and is south facing (Donovan 1965). The second felsic unit occurs in Heenan Township, on the south limb of the syncline. It lies south of the komatiites on this south fold limb (Goodwin 1965). The east-striking, axial trace of the Marion River anticline lies 3 km to the south of this felsic unit (Goodwin 1965).

A narrow, discontinuous, east-southeast-striking, komatiitic unit (less than 2 km), having a high magnetic signature (OGS 1982), occurs close to the southern boundary of the Halcrow–Swayze assemblage. If correlative with the northern komatiites, this distribution suggests that the komatiite-bearing unit has been folded, on a larger scale, about a west-northwest-striking, west-closing anticline which may correlate with the Marion River anticline.

A U-Pb zircon age of 2697 Ma for felsic metavolcanic rocks of the Raney–Newton assemblage and a Sm-Nd age of 2826±64 Ma for tholeiitic and komatiitic flows of the Halcrow–Swayze assemblage (Cattell et al. 1984) may indicate a substantial age difference between the 2 assemblages. If the age data are correct, metavolcanic rocks of the Halcrow–Swayze assemblage are juxtaposed against the Raney–Newton assemblage.

Marion Assemblage

The Marion assemblage (MN) is a moderately to steeply dipping, wedge-shaped, east-facing assemblage (*see* Figure 11.7) that consists of intermediate to felsic flows and tuff breccia metamorphosed to greenschist and amphibolite facies (Goodwin 1965). Quartz- and feldspar-phyric rhyolitic and dacitic, massive flows are interlayered with poorly sorted pyroclastic units and minor andesitic flows and tuffs (Goodwin 1965). Coarser rhyolitic tuff and breccia occur adjacent to the Woman River iron formation, which caps the assemblage (Goodwin 1965). The eastern part of the assemblage has been intruded by a massive, hornblende-biotite granodiorite phase of the Kenogamissi Batholith and younger bodies of diorite and gabbro (Goodwin 1965).

Immediately adjacent to the Woman River iron formation, felsic metavolcanic rocks have been replaced by magnetite, pyrite and siderite (Goodwin 1965). The iron formation is steeply dipping and consists of interlayered chert, jasper, magnetite, sulphide (pyrite-pyrrhotite) and carbonaceous and siderite mesobands (Goodwin 1965). Pillow basalts in the Halcrow–Swayze assemblage, which immediately overlie the iron formation, face to the east and north (Goodwin 1965). Discontinuous units of iron formation are present along the southwestern side of the assemblage (J.C. Ireland, Porcupine Resident Geologist's office, personal communication, 1990).

The metavolcanic assemblage is folded about the west-striking Marion River anticline having a well-developed axial planar foliation (Goodwin 1965). The northwestern contact between the iron formation and mafic flows of the Halcrow–Swayze assemblage to the west is interpreted to be conformable (Goodwin 1965). However, along the southern contact with the Halcrow–Swayze assemblage, the abrupt change in strike and the asymmetrical fold limbs may indicate that this may be a tectonic contact.

Garnet–Tooms Assemblage

The Garnet–Tooms assemblage (GT) is a greenschist to amphibolite facies assemblage (*see* Figure 11.7) which is delimited on the north by the Ridout assemblage and to the south by a unit of oxide-facies iron formation, which caps the Hong Kong assemblage (*see* Hong Kong Assemblage). To the west, the assemblage is truncated by a north-striking fault, which juxtaposes the metavolcanic rocks against gneissic granodiorite. The assemblage consists of tholeiitic basalt, intermediate to felsic, calc-alkalic flows and fragmental rocks, and komatiitic flows and fragmental rock, interlayered with oxide-facies iron formation (Meen 1944; Siragusa 1987; J.C. Ireland, Porcupine Resident Geologist's office, personal communication, 1990). Multiple, thin (less than 10 m) units of steeply dipping, oxide-facies iron formation occur at the northern or upper part of the assemblage. Iron formation is also found elsewhere in the assemblage (Meen 1944). Coarser-grained dioritic and gabbroic phases, representing intrusions or thick flows, also occur within the basaltic unit (Moorhouse 1951). Several east-striking, east-closing folds occur in the western part of the assemblage (J.C. Ireland, Porcupine Resident Geologist's office, personal communication, 1990).

The central part of the assemblage thins to 2 km in width in map view and dips and faces to the north (Meen 1944). Massive and pillowed tholeiitic basalts constitute the southern half of this part of the assemblage (Meen 1944; Siragusa 1987), the basal unit of which is massive, feldspar-megacrystic rock ("leopard rock"; Meen 1944). Gabbroic and dioritic sills are also associated with the tholeiitic metavolcanic rocks (Siragusa 1987). The upper and central part of the assemblage consists mainly of calc-alkalic, feldspar-porphyritic (less than 3 mm) basaltic and andesitic flows (Meen 1944; Siragusa 1987).

The eastern part of the assemblage consists of pillowed, tholeiitic basalt interlayered with minor oxide-facies iron formation and intermediate to felsic metavolcanic rock (Laird 1936a; Moorhouse 1951; Siragusa 1987). A felsic unit, exposed south of Opeepeesway Lake, consists of banded, slaty tuff (sheared?) and laminated (sheared?) quartz porphyry (Moorhouse 1951) and is associated with chert-rich iron formation. The mineralogical similarity with the chert-rich iron formation which caps the Hong Kong assemblage lying to the south (*see* Hong Kong Assemblage) may indicate that the Garnet–Tooms assemblage can be further subdivided. Along the southern margin of the assemblage, adjacent to a gneissic external granodiorite, the metavolcanic rocks are sheared, locally gneissic and have been contact metamorphosed to amphibolite facies (Laird 1936a; Moorhouse 1951; Siragusa 1987). A well-developed penetrative, southeast-striking foliation has obscured most primary textures. Flows adjacent to the granodiorite intrusion face north, although south-facing flows occur adjacent to the contact with the Ridout assemblage (Moorhouse 1951).

To the north, the contact with the Ridout assemblage is locally sheared (Furse 1932). To the south, the contact with the metavolcanic assemblage (*see* Hong Kong Assemblage) is also locally sheared (J.C. Ireland, Porcupine Resident Geologist's office, personal communication, 1990).

Associated with some cherty iron formations are base and precious metal veins and replacement zones (e.g., Thurston et al. 1977).

Hong Kong Assemblage

The Hong Kong assemblage (HK) crops out along the southern edge of the Swayze area and extends southeast to the Biscotasing region (*see* Figure 11.7), where amphibolite facies gneissic rocks crop out (Thurston et al. 1977). The assemblage consists of steeply dipping, tholeiitic basalts, interlayered with minor, thin (less than 10 m) units of clastic metasedimentary rock and oxide-facies iron formation (Meen 1944). A unit of chert-jasper-magnetite iron formation occurs along the northern limit of the assemblage (Meen 1944), the trace of which outlines a series of tight folds in the central part of the assemblage having a northerly axial trace (Meen 1944). The trace of the iron formation cap is consistent with a north-striking anticlinal fold; however, the dominant penetrative foliation dips steeply and strikes easterly, parallel to the contact with the external granitoids and is not axial planar to the folds. The contact with the Garnet–Tooms assemblage is presumed to be conformable, although it is locally sheared (Meen 1944).

Assemblage Correlation

The Hanrahan and Marion assemblages consist predominantly of felsic metavolcanic rocks, which are cut by numerous mafic and ultramafic sills and dikes; each unit is capped by oxide-facies iron formation (Goodwin 1965; Milne 1972). These assemblages occur adjacent to the western margin of the Kenogamissi Batholith. Iron formation extends along the northern edge of the Kenogamissi Batholith, and may correlate with iron formation units which cap the 2727±1.5 million-year-old (Corfu et al. 1989) Bartlett assemblage on the eastern side of the Kenogamissi Batholith. The close association of these 3 metavolcanic assemblages with the margin of the Kenogamissi Batholith, and their lithological similarity, including the iron formation "caps", suggests that these assemblages may be correlative, possibly representing some of the oldest metavolcanic rocks in the southern part of the Abitibi greenstone belt. The Hong Kong assemblage displays a similar spatial relationship to external batholiths, and is also capped by oxide-facies iron formation; hence, this assemblage may also be correlative with the Marion, Hanrahan and Eldorado assemblages. However, the Hong Kong assemblage consists predominantly of mafic metavolcanic rocks, hence, stratigraphic correlation is cautioned.

BATCHAWANA GREENSTONE BELT

The Batchawana greenstone belt (*see* Figures 11.2 and 11.28) is subdivided into the metavolcanic Griffin and Dismal assemblages, and the metasedimentary Wart assemblage (*see* Figure 11.28; Giblin and Armburst 1973; Grunsky 1980, 1981; Corfu and Grunsky 1987). The greenstone belt is (Corfu and Grunsky 1987): 1) gradational into migmatites of the Ramsey Gneiss Domain to the northwest and east; 2) intruded to the south by the Algoma Plutonic Domain, which includes massive tonalite and granodiorite intrusions and some gneissic units; and 3) separated from the Chapleau Gneiss Domain migmatites and syn- to post-massive granodiorite and granite (Card 1979) by the Montreal River fault.

The western part of the greenstone belt is overlain by Keweenawan metavolcanic rocks, Upper Keweenawan metasedimentary rocks and the Mesoproterozoic (Cannon et al. 1989 and references therein) Jacobsville Formation, which was previously assigned a Lower and Middle Cambrian age by Giblin and Armburst (1973). Metamorphic grade in the belt varies from greenschist in the central areas to amphibolite facies; locally, anatectic migmatites exist near the plutonic-gneiss domains (Grunsky 1980, 1981).

Synvolcanic, 2716 million-year-old (Corfu and Grunsky 1987) external tonalite plutons were emplaced into the southern part of the greenstone belt (*see* Griffin Assemblage) and into the Algoma Plutonic Domain. Isoclinal folding and metamorphism of greenstone belt units occurred between 2677 and 2668 Ma (Corfu and Grunsky 1987), coincident with the emplacement of structurally concordant, syntectonic to late tectonic diorite, tonalite, and granodiorite plutons into the Ramsey Gneiss Domain, and the emplacement of several small, posttectonic, internal plutons (e.g., Grey Owl Lake granite, 2673±5 Ma; Griffin granodiorite, 2674±3 Ma; Corfu and Grunsky 1987) into the greenstone belt. Contact metamorphic aureoles, which are related to the late internal intrusions, are superimposed upon the regional metamorphic pattern (Corfu and Grunsky 1987). The presence of slightly older xenocrystic zircons (greater than 2680 Ma) in some granodiorite intrusions, which were emplaced into the Ramsey Gneiss Domain, indicates that some of the magmas may have originated, in part, by partial melting of crustal material (Corfu and Grunsky 1987).

Diabase dikes of the northwest-striking Sudbury and northeast-striking Preissac dike swarm (Fahrig and West 1986) cut all Precambrian rocks (*see* Osmani, this volume).

Few metal resources are known from the Batchawana greenstone belt. Most mineral deposits in the belt are located in the western part of the greenstone belt. Both Archean and Proterozoic copper mineralization is present. Proterozoic copper mineralization consists of carbonate-quartz-barite-chalcopyrite-pyrite-molybdenite within breccia pipes cutting Archean metavolcanic and granitoid rocks (Shklanka 1969, p.48). A second type of Proterozoic copper mineralization consists of disseminated and vein chalcocite-carbonate, localized within Middle Keweenawan volcanic and sedimentary rocks (Shklanka 1969, p.60).

Griffin Assemblage

The 2729 to 2716 million-year-old (Corfu and Grunsky 1987) Griffin assemblage (GF) is a steeply dipping, east-trending assemblage of iron- and magnesium-rich tholeiitic basalt, interlayered with minor felsic tuffs and metasedimentary rocks (Giblin and Armburst 1973; Grunsky 1980, 1981). Within the mafic metavolcanic units, pillowed, massive, and fragmental volcanic facies are present (Giblin and Armburst 1973). The assemblage is mainly north facing, although some south-facing strata exist (Giblin and Armburst 1973).

Within the Griffin assemblage, a steeply dipping, east-striking penetrative foliation is developed (Giblin and Armburst 1973). Reversals in facing directions suggest strata are folded, but paucity of data precludes delineation of regional folds.

The contact with the Wart metasedimentary assemblage to the east is, in part, a shear zone. The contact with the Ramsey Gneiss Domain to the north corresponds with north- to northeast-striking faults (*see* Figure 11.28).

Copper occurrences of possible Archean age occur in the Griffin assemblage, where disseminated, fracture and vein chalcopyrite, pyrite, molybdenite and quartz-carbonate mineralization cuts Archean quartz and feldspar porphyry dikes and spatially associated mafic metavolcanic rocks. This style of copper mineralization is similar to Proterozoic copper occurrences.

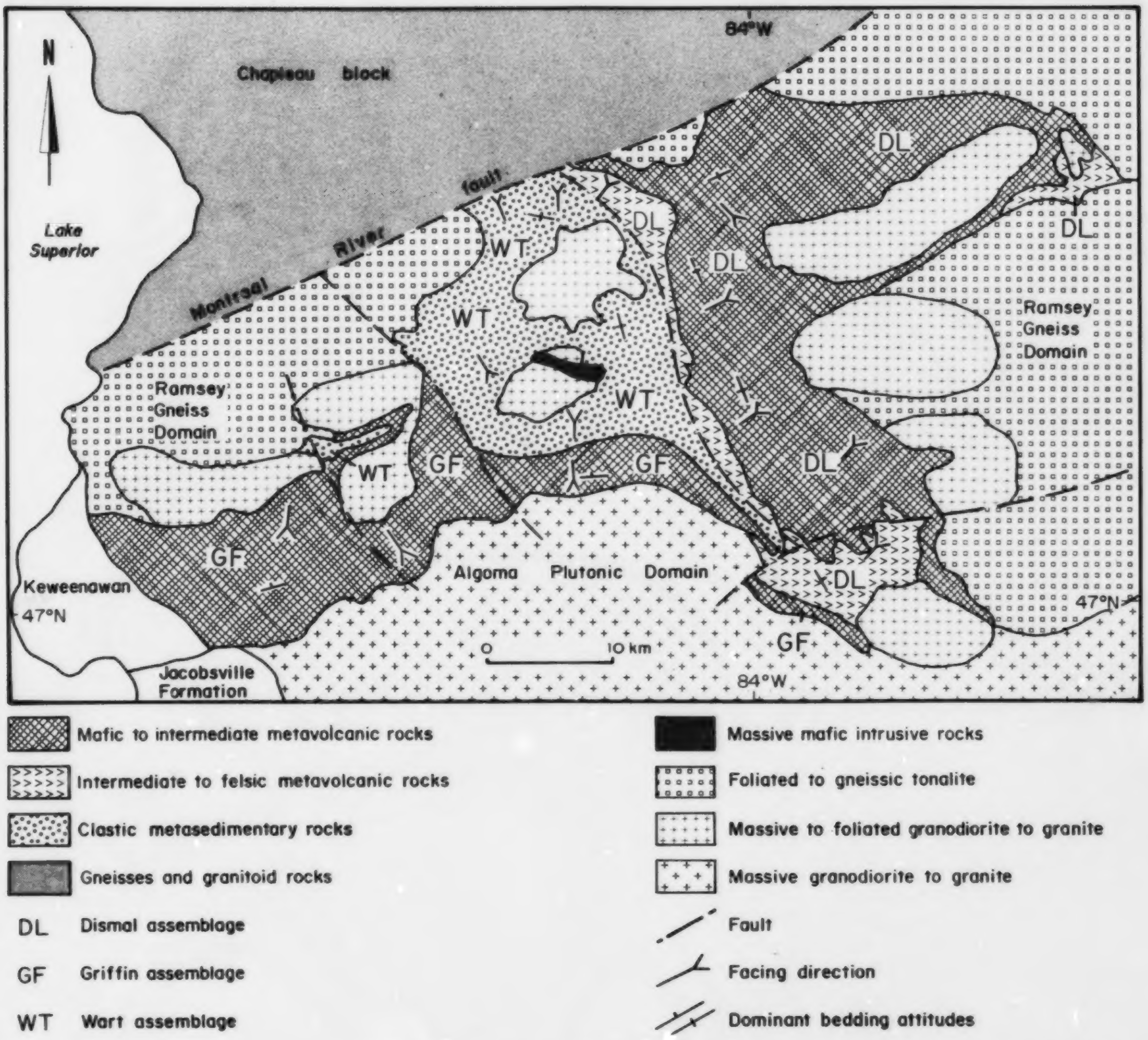

N
84°W
Chapleau block
Lake Superior
Montreal River fault
DL
WT
GF
Ramsey Gneiss Domain
Algoma Plutonic Domain
Keweenawan
47°N
Jacobsville Formation
0
10 km
84°W
47°N
Mafic to intermediate metavolcanic rocks
Intermediate to felsic metavolcanic rocks
Clastic metasedimentary rocks
Gneisses and granitoid rocks
DL Dismal assemblage
GF Griffin assemblage
WT Wart assemblage
Massive mafic intrusive rocks
Foliated to gneissic tonalite
Massive to foliated granodiorite to granite
Massive granodiorite to granite
Fault
Facing direction
Dominant bedding attitudes

Figure 11.28. Generalized geology of the Batchawana greenstone belt area (*modified from* Corfu and Grunsky 1987). Unless otherwise indicated, all rocks are Archean in age. The Ramsey Gneiss Domain and Algoma Plutonic Domain, collectively, comprise the Ramsey–Algoma granitoid complex. The Jacobsville Formation represents upper Proterozoic sandstone and siltstone deposits. Keweenawan rocks consist principally of Proterozoic basalt. *See* Figure 11.2 for location of area. The Montreal River fault separates the Chapleau block (gneisses) from units to the southeast and, in this area, marks the boundary between the Wawa Subprovince and the Abitibi Subprovince.

Dismal Assemblage

The westward-facing, 2711 to 2698 million-year-old (Corfu and Grunsky 1987) Dismal assemblage (DL; *see* Figure 11.28) consists of both tholeiitic and calc-alkalic pillowed flows interlayered with minor breccias, tuffaceous units, clastic sedimentary rocks and komatiitic basalt (Grunsky 1980). Pillowed, amygdaloidal, variolitic, and plagioclase-megacrystic, iron-rich, tholeiitic basalt is abundant in the northwestern, northern and eastern parts of the assemblage (Grunsky 1980). Plagioclase- and quartz-porphyritic calc-alkalic andesitic, dacitic and rhyolitic flows, pyroclastic breccia, lapillistone and crystal and ash tuff predominate in the southeastern part of the assemblage (Grunsky 1980). Oxide and sulphide facies iron formation occurs in the eastern part of the area, interlayered with both tholeiitic and calc-alkalic volcanic rocks (Grunsky 1980).

Regional foliation has a northwest strike (Grunsky 1980). An east- to northeast-trending syncline, located in the northeastern part of the assemblage, appears to merge into a larger southeast-to south-trending synclinorium, centred in the Wart assemblage.

Wart Assemblage

The Wart assemblage (WT) consists of south-facing metasedimentary rocks including wacke, arkose and conglomerate (Grunsky 1980, 1981). At its base, the Wart assemblage is conformably interlayered with the Dismal assemblage (Grunsky 1980, 1981). In the Wart assemblage, regional foliation strikes northwest, except adjacent to the Griffin assemblage, and in the northern areas, where a more easterly foliation trend is developed. Along the north contact, bedding is rotated in an anticlockwise and clockwise sense in the metasedimentary and metavolcanic assemblages, respectively. Thus bedding parallels the contact between the metavolcanic and metasedimentary rocks (Grunsky 1981). This is consistent with the presence of a northwest-striking dextral shear zone, localized along the contact between the 2 assemblages.

BENNY, HUTTON, PARKIN AND TEMAGAMI GREENSTONE BELTS

Several narrow greenstone belts crop out within and to the east of the Ramsey–Algoma granitoid complex, north of the Sudbury Basin. These greenstone belts are described together for convenience; however, they may represent parts of a formerly more extensive greenstone belt.

BENNY GREENSTONE BELT

The greenschist to amphibolite facies, Benny greenstone belt (*see* Figures 11.2 and 11.29; Card and Innes 1976a, 1976b, 1976c, 1981; Guthrie 1980) is approximately 35 km long and 5 km wide and is bounded by gneissic, foliated and massive Archean granitoid rocks of the Ramsay–Algoma granitoid complex. In the eastern and northern parts of the belt, the metavolcanic rocks pass gradationally into migmatitic rocks consisting of variable proportions of granitic material and large xenoliths of mafic and intermediate metavolcanic schists (Card and Innes 1981; Guthrie 1980). Foliated, massive and porphyritic granite and granodiorite-trondhjemite intrusions cut both the gneissic rocks and the greenstone belt; younger intrusions are quartz monzonite (Card and Innes 1981). The contacts between the greenstone belt and the adjacent foliated granodiorite-trondhjemite plutonic rocks are generally sheared (Card and Innes 1981). The east-central and northwestern parts of the belt are informally subdivided into the Geneva and Bluewater assemblages, respectively (*see* Figure 11.29).

A regionally developed, penetrative, east-striking, flattening fabric is approximately parallel to stratigraphic contacts. The dip of the foliation is towards the core of the belt. Mineral and stretching lineations generally plunge steeply downdip (Card and Innes 1981). Large-scale folds are not recognized, although undulations in stratigraphic trends suggest gentle folding (Guthrie 1980). Foliation and gneissosity in the surrounding migmatitic and plutonic rocks generally strike eastward, parallel to the greenstone belt (Guthrie 1980).

Geneva Assemblage

The Geneva assemblage (GE) consists of calc-alkalic andesitic and dacitic-rhyolitic flows, pyroclastic tuffs and breccias, and cherty and calcareous metasedimentary rocks, graphitic shale and minor turbidite (Guthrie 1980). In the central part of the assemblage, several cyclically repeated volcanic sequences, consisting of minor tholeiitic (basaltic) and calc-alkalic intermediate and felsic units, occur interlayered with sulphidic and carbonaceous metasedimentary units (Card and Innes 1981). Coarse-grained pyroclastic andesitic and rhyolitic rocks are more common in the eastern part of the assemblage (Card and Innes 1981).

Minor units of volcaniclastic, resedimented and sulphide-rich chemical metasedimentary rocks occur (Guthrie 1980). Thick, laterally continuous accumulations of oxide-facies (chert-magnetite) iron formation are not recognized in the Benny greenstone belt (Card and Innes 1981; Guthrie 1980).

Known economic mineralization consists of the Geneva Lake Mine (Card and Innes 1981), a volcanic associated, zinc- and lead-rich massive sulphide deposit. This stratabound mineralization occurs within calc-alkalic rocks near the top of the assemblage (Hawley 1948; Guthrie 1980; Card and Innes 1981). Another base metal occurrence lies 14 km to the west near the village of Stralak, and occurs in a similar stratigraphic setting (Card and Innes 1976a, 1976b, 1981).

Bluewater Assemblage

The Bluewater assemblage (BW; Guthrie 1980) is composed mainly of iron-rich, tholeiitic basalt flows with minor intercalated, calc-alkalic intermediate and felsic volcaniclastic and tuffaceous metasedimentary rocks (Card and Innes 1981; Guthrie 1980). Gabbroic intrusions occur preferentially within this assemblage (Card and Innes 1981). Most of the flows are massive, although pillowed,

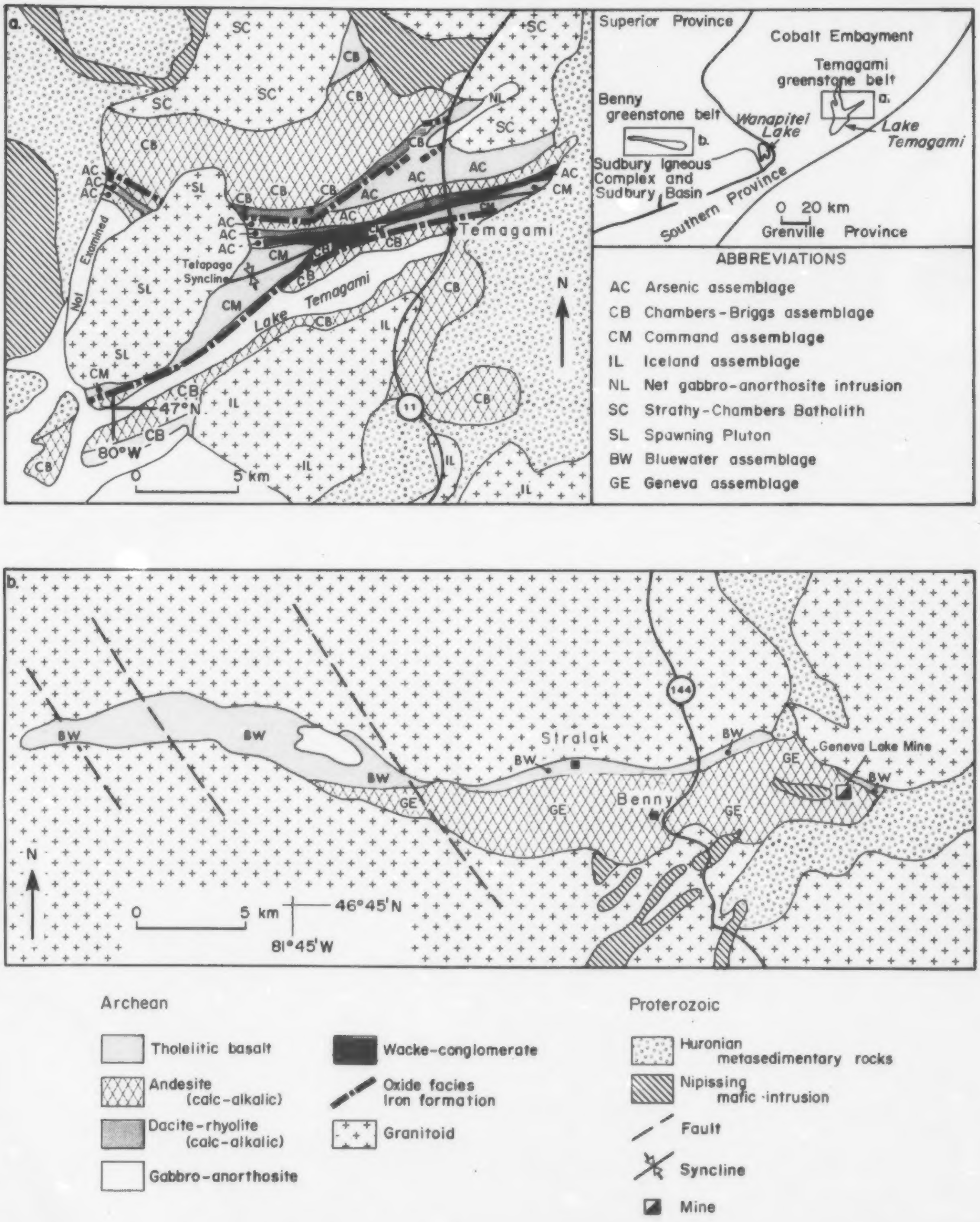

a.
SC
CB
NL
AC
CM
SL
IL
Not Examined
Tetapaga Syncline
Lake Temagami
Temagami
47°N
80°W
0
5 km
N
11
Superior Province
Cobalt Embayment
Temagami greenstone belt
Benny greenstone belt
Wanapitei Lake
Lake Temagami
Sudbury Igneous Complex and Sudbury Basin
Southern Province
0 20 km
Grenville Province
ABBREVIATIONS
AC Arsenic assemblage
CB Chambers-Briggs assemblage
CM Command assemblage
IL Iceland assemblage
NL Net gabbro-anorthosite intrusion
SC Strathy-Chambers Batholith
SL Spawning Pluton
BW Bluewater assemblage
GE Geneva assemblage
b.
144
BW
GE
Stralak
Benny
Geneva Lake Mine
N
0
5 km
46°45'N
81°45'W
Archean
Tholeiitic basalt
Andesite (calc-alkalic)
Dacite-rhyolite (calc-alkalic)
Gabbro-anorthosite
Wacke-conglomerate
Oxide facies iron formation
Granitoid
Proterozoic
Huronian metasedimentary rocks
Nipissing mafic intrusion
Fault
Syncline
Mine

Figure 11.29. Generalized geology of the: **a)** Temagami Greenstone Belt area (*modified from* Bennett 1978; Fyon and Cole 1989); and **b)** Benny greenstone belt area (*after* Guthrie 1980; Card and Innes 1981). Adjacent granitoid rocks belong to the Ramsey–Algoma granitoid complex. Highways 144 and 11 are denoted by circled numbers.

porphyritic and amygdaloidal varieties are present. The presence and distribution of pillowed units is difficult to determine because of the generally high state of strain (Card and Innes 1981). Based on northward facings throughout the belt, Guthrie (1980, p.75) concluded that the Bluewater assemblage conformably overlies the Geneva assemblage. Alternatively, Card and Innes (1981) assigned a low confidence to the facing directions due to the high-strain state but postulated a south-facing succession. All units dip towards the centre of the belt, away from the margins with the Ramsay–Algoma granitoid complex (Card and Innes 1981). Ultramafic and other magnesium-rich flows are not reported.

HUTTON AND ROBERTS GREENSTONE BELTS

Several small greenstone belts are exposed north of the Sudbury Basin, between the Benny and Temagami greenstone belts, in the Hutton–Roberts area (Meyn 1970, 1971). Within these greenstone belts, massive, pillowed and layered (sheared?) basaltic tholeiitic metavolcanic rocks predominate (Meyn 1970). Intermediate to felsic pyroclastic and effusive metavolcanic and clastic metasedimentary rocks are not abundant (Meyn 1970). Felsic schists, some of which contain lenses of conglomerate, may represent either clastic metasedimentary rocks or sheared intermediate to felsic metavolcanic rocks (Meyn 1971). Chemical metasedimentary rocks, consisting mainly of laterally continuous chert-magnetite-amphibole iron formation, occur interlayered with mafic flows (Meyn 1970), and were mined as an iron deposit at the Moose Mountain Mine in Hutton Township.

Metamorphic grade of the greenstone assemblages is lower greenschist to almandine-amphibolite facies (Meyn 1970, 1971). A steeply dipping, penetrative foliation in the metavolcanic rocks strikes southeast, subparallel to general stratigraphic contacts (Meyn 1970).

Granodiorite gneiss is the oldest known granitoid; it contains inclusions of quartz-biotite and quartz-hornblende schists, possibly representing greenstone xenoliths. If the xenoliths represent supracrustal rocks derived from the greenstone belt, then they indicate a postvolcanic emplacement of the gneissic rocks. Massive to foliated granite and granodiorite intrusions cut both the metavolcanic and gneissic rocks (Meyn 1970). Small bodies of massive, feldspar-megacrystic granite locally occur (Meyn 1970, 1971). The granitoid rocks are, in turn, intruded by altered (Archean) mafic dikes.

The Parkin offset dike, related to the Proterozoic Sudbury Igneous Complex, is host to copper-nickel sulphide mineralization at the Jonsmith Mine (Meyn 1970). Placer gold is concentrated in the Pleistocene glaciofluvial gravels of the Vermilion River (Meyn 1970), but no economic concentrations have been located.

TEMAGAMI GREENSTONE BELT

Metavolcanic and metasedimentary rocks in the Temagami Greenstone Belt (*see* Figures 11.2 and 11.29) strike east to northeast and have been folded about the east-striking Tetapaga syncline (Bennett 1978). Coeval with the extrusion of the volcanic rocks was the emplacement of: 1) layered pyroxenite-gabbro-anorthosite sills of tholeiitic affinity (Fyon and O'Donnell 1987; Fyon et al. 1988); 2) a layered dunite-peridotite-gabbro plug of komatiitic affinity (James and Hawke 1984; Good 1989); and 3) diorite and quartz-diorite sills of calc-alkalic affinity (Bennett 1978). The layered, gabbro-anorthosite bodies occur spatially associated with coarse-grained (1 to 5 cm), feldspar-megacrystic, iron-rich, tholeiitic basalt flows (Fyon and O'Donnell 1987; Fyon et al. 1988). Intrusive into the belt is the massive to foliated tonalite-trondhjemite Strathy–Chambers Batholith and the trondhjemite-granodiorite Iceland Batholith. A massive, microcline-megacrystic granite stock (Spawning Lake) cuts across all stratigraphic and structural trends. Lamprophyre and pyroxenite dikes cut all Archean units, but do not cut the Proterozoic units (Bennett 1978).

Greenschist facies metamorphic conditions prevailed, except adjacent to the late granitoids where amphibolite facies conditions were attained. The development of penetrative foliation varies from indistinct to intense. Early foliations have an easterly strike which is parallel to stratigraphic contacts, the axial planar foliation of the Tetapaga syncline, and the margins of the Iceland and Strathy–Chambers batholiths. Parallel to this earlier foliation are the Northeast Arm deformation zone and the Link Lake shear zone (Fyon and O'Donnell 1987; Fyon et al. 1988). The east-striking foliation, the Tetapaga syncline and the paired Northeast Arm deformation zone and Link Lake shear zone are interpreted to have developed contemporaneously (Fyon and O'Donnell 1987). The later, east- to northeast-striking Net–Vermilion deformation zone and Tasse shear zone are characterized by an oblique, sinistral component of slip (Bennett 1978; Fyon and O'Donnell 1987; Fyon et al. 1988). The related shear fabric cuts parts of the Strathy–Chambers Batholith (Fyon and O'Donnell 1987) and deflects the earlier east-striking foliation (Fyon and O'Donnell 1987; Fyon et al. 1988). A later north-striking foliation folds the earlier east-striking foliation and the northeast-striking shear foliation (Fyon et al. 1988).

Arsenopyrite, pyrrhotite, chalcopyrite, sphalerite, galena and auriferous pyrite concentrations, in both quartz vein and impregnation habits, are localized within: 1) a northeast-striking segment of the Net–Vermilion deformation zone south of the Strathy–Chambers Batholith; and 2) shear zones perpendicular to the southeastern lobe of the Strathy–Chambers Batholith in Strathy Township (Bennett 1978; Fyon and Crocket 1986; Fyon et al. 1988).

Chambers–Briggs Assemblage

The 2737 million-year-old (Bowins and Heaman, in press) Chambers–Briggs assemblage (CB; *see* Figure 11.29) consists of massive, pillowed and coarse-grained feldspar-megacrystic tholeiitic basalts at the base, overlain by calc-alkalic, effusive and pyroclastic andesite flows, coarse-grained (less than 100 cm), resedimented, andesitic debris deposits, and dacite and effusive rhyolite flows and subaqueous pyroclastic flows (Bennett 1978; Fyon and O'Donnell 1987; Born 1988; Fyon et al. 1988). The resedimented debris-flow deposits occur preferentially in the northwestern part of the greenstone belt. The assemblage is capped by units of pyrite- pyrrhotite and chert-magnetite-hematite iron formation (Bennett 1978; Fyon et al. 1988; Fyon and Cole 1989). Stratigraphic units on the north limb of the Tetapaga syncline have steep dips and face to the south. Stratigraphically equivalent rock units on the south limb also have steep dips, but face to the north (Bennett 1978).

Two discrete iron formations occur at the top of the Chambers–Briggs assemblage. A thinner (less than 25 m) unit of chert, pyrite and pyrrhotite (sulphide-facies iron formation) lies stratigraphically below a thicker unit (100 m) of chert-magnetite-hematite oxide-facies iron formation (Bennett 1978). On the south limb of the Tetapaga syncline, the oxide-facies iron formation is interlayered with a turbiditic metasedimentary unit (Fyon and Crocket 1986). On the north limb of the Tetapaga syncline, a unit of ultramafic fragmental rock, magnesium-rich (12 to 20 weight % MgO) flows and coarse-grained heterolithic fragmental rocks of uncertain origin, overlie the oxide-facies iron formation (Bennett 1978; Fyon et al. 1988). The chert-magnetite-hematite-chlorite iron formations have been the source of iron ore extracted at the Sherman iron mine.

Polymetallic, vein, disseminated and massive base metal sulphide mineralization, enriched in copper, gold, cobalt, platinum and silver, was extracted at the Temagami copper mine (e.g., Colvine 1974). The mineralization is localized along the contact between a diorite intrusion and intermediate, calc-alkalic metavolcanic rock near the top of the Chambers–Briggs assemblage, on the south limb of the Tetapaga syncline (Bennett 1978).

Arsenic Assemblage

The wedge-shaped Arsenic assemblage (AC; *see* Figure 11.29; Fyon and Crocket 1986) consists of steeply dipping, south-facing, iron-rich, massive, pillowed, coarse-grained, feldspar-megacrystic, and variolitic tholeiitic basalt and andesitic to rhyolitic effusive flows and fragmental rocks (Fyon et al. 1988). Vesicularity of flow units increases from zero at the base of the assemblage to consistently greater than 10% at the top of the assemblage (Fyon and Crocket 1986). The intermediate to felsic metavolcanic rocks consist of andesitic effusive flows, commonly with flow top breccias and rhyolitic fragmental rocks interpreted to have been deposited as subaqueous pyroclastic flows (Figure 11.30; Bennett 1978; Fyon and Crocket 1986). The rhyolitic rocks have some geochemical characteristics of the high-silica rhyolites (Lesher et al. 1986) associated with volcanogenic massive base metal sulphide mineralization (Fyon and Crocket 1986; Fyon and Cole 1989). Unlike the Chambers–Briggs assemblage, resedimented andesitic debris flows are not abundant. The intermediate to felsic metavolcanic rocks are overlain conformably by turbiditic metasedimentary rocks. Regionally continuous sulphide and oxide-facies iron formation units are not observed in the Arsenic assemblage, although thin, discontinuous, sulphide-rich interflow units occur at the top of the assemblage.

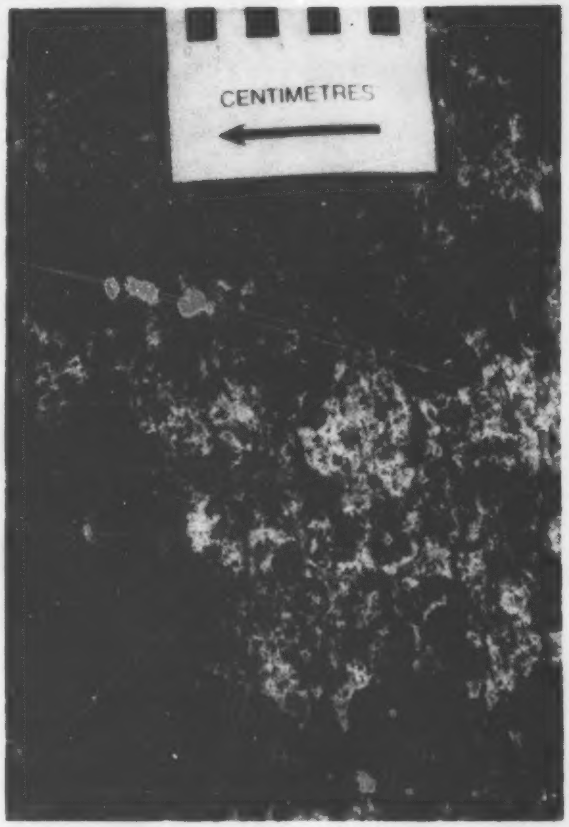


Figure 11.30. Subaqueous, pyroclastic, high-silica rhyolite with dark, wisp-like pumiceous fiamme, top of Arsenic assemblage, Temagami greenstone belt.

The contact between the Arsenic and Chambers–Briggs assemblages is the locus of the oblique, sinistral Net–Vermilion deformation zone (Fyon and O'Donnell 1987; Fyon et al. 1988; Fyon and Cole 1989). Along the western part of this contact, the unit of tholeiitic basalt and part of the andesitic unit belonging to the Arsenic assemblage have been structurally removed (Fyon et al. 1988). This stratigraphic relationship and the geometry of the shear zone are consistent with an early thrust history (Fyon et al. 1988). The nature of the contact between the clastic metasedimentary rocks of the Arsenic assemblage and the metavolcanic rocks of the Command assemblage (*see* Command Assemblage) to the south is unclear (Fyon and Cole 1989).

Command Assemblage

The Command assemblage (CM; *see* Figure 11.29) consists of massive and pillowed, iron-rich tholeiitic basalt. The assemblage occurs only in the core of the Tetapaga syncline

(Fyon and Cole 1989). The strike and dip of flows within this assemblage are poorly constrained because of the absence of distinctive marker units (Bennett 1978). Also, contact relationships between this and adjacent assemblages are not well constrained.

INTRUSIONS OF THE WESTERN ABITIBI SUBPROVINCE

Granitoid intrusions range from tonalite and diorite through to alkali-feldspar syenite (*see* Map 2543, map case). These intrusions form isolated bodies within the greenstone belts (e.g., Otto Stock), intermediate-sized batholiths within the greenstone belts (e.g., Round Lake and Lake Abitibi batholiths) and large complexes that bound or separate greenstone belts (e.g., Ramsey–Algoma granitoid complex). Additionally, there are numerous dikes ranging from aplitic felsite to quartz-feldspar porphyry and lamprophyre dikes. Mafic to ultramafic intrusions (e.g., Kamiskotia Gabbroic Complex, Ghost Range Complex) are also volumetrically significant.

Granitoid Intrusions

REGIONAL SPACE-TIME-COMPOSITION RELATIONSHIPS

Based on high-precision uranium-lead zircon geochronology and crosscutting relationships (e.g., Marmont and Corfu 1989; Corfu et al. 1989), the following pattern of granitoid magmatism in the western Abitibi Subprovince appears to be emerging. Prior to 2700 Ma, large trondhjemite-tonalite-granodiorite bodies were emplaced, including the Round Lake batholith, much of the Ramsey–Algoma granitoid complex, the Lake Abitibi Batholith and a small part of the southwestern corner of the Watabeag Batholith. The granodiorite phases of these large bodies are generally younger than the foliated to gneissic tonalitic phases. Although some ~2700 million-year-old metavolcanic units are subparallel to some batholith contacts (e.g., Skead assemblage and Round Lake batholith) and thus, were likely deformed by emplacement of the batholith, although not necessarily in the magmatic state, no tonalite-trondhjemite batholiths are known to intrude, as magmatic phases, the 2700 million-year-old and younger metavolcanic units. Between ~2700 and 2675 Ma, following crystallization of the large batholithic complexes, discrete granodiorite-granite plutons were emplaced within the southern Abitibi greenstone belt and at greenstone-batholith margins (e.g., northwestern Batchawana greenstone belt). During this period, a large number of feldspar ± quartz porphyry dikes were emplaced (e.g., Timmins area). Following this period, smaller discrete alkalic plutons were emplaced near, but not necessarily confined to, the regional scale shear zones (Porcupine–Destor and Larder–Cadillac) between ~2685 and 2675 Ma. These alkalic plutonic rocks may be related to the coeval alkalic metavolcanic rocks preserved within the Timiskaming assemblage (*see* Timiskaming Assemblage). During and subsequent to the emplacement of the alkalic plutons, lamprophyre dikes were emplaced.

RAMSEY–ALGOMA GRANITOID COMPLEX

The Ramsey–Algoma granitoid complex (*see* Figure 11.2) consists of the Ramsey Gneiss Domain to the north, the Algoma and Levack Gneiss domains to the south and the central Algoma Plutonic Domain. This complex is bounded to the: 1) south by the Paleoproterozoic Huronian Supergroup of the Southern Province; 2) east by faults that bound the Cobalt Embayment; and 3) west by gneisses of the Chapleau Gneiss Domain of Card (1979). Little information is available on this part of the Superior Province and its relationships to other elements of the central Superior Province are unknown. The following summary is taken directly from Card (1979, p.88):

> The Ramsey Gneiss Domain, which links the Abitibi–Swayze and Batchewana metavolcanic domains, contains granodioritic and agmatitic granitic gneiss and foliated to massive granodiorite-granite plutons. Within the Ramsey Gneiss Domain there are abundant xenoliths and enclaves of supracrustal rocks, mainly of volcanic origin, and the largest of these has been distinguished as the Biscotasing metavolcanic domain. The Algoma Gneiss Domain, which surrounds and underlies the Proterozoic Huronian rocks of the Southern Province, consists of granitic, granodioritic and granitic gneiss with numerous greenstone enclaves and massive to foliated granite, granodiorite, and syenite intrusions. The Levack Gneiss Domain forms a narrow belt marginal to the North Range of the Sudbury Nickel Irruptive and consists of cataclastically foliated, high-rank metamorphic (amphibolite, granulite facies) granitic-dioritic gneiss with mafic xenoliths and enclaves of probable metavolcanic origin. The Levack Gneiss Domain represents the north limb of a major domal structure upon which the Sudbury Structure was superimposed (Card and Hutchinson, 1972). The Algoma Plutonic Domain constitutes one of the largest areas of massive felsic plutonic rocks in the southern Superior Province. This large batholithic complex consists mainly of coarse, porphyritic, leucocratic granite or quartz monzonite with subordinate granodiorite. The complex was emplaced as a series of magmatic intrusions at intermediate crustal levels following deformation and regional metamorphism of the gneiss and greenstone.

One unit from the Levack Gneiss Domain has been dated at 2711±7 Ma (Krogh et al. 1984). One unit of the Algoma Gneiss Domain has been dated at 2716±2 Ma (Corfu and Grunsky 1987).

EARLY BATHOLITHS

The early batholiths are mostly members of the trondhjemite-tonalite-granodiorite suite.

Round Lake Batholith

The Round Lake batholith (*see* Figure 11.2) is a 40 by 80 km, elliptical, east-northeast-trending body with a vertical extent of approximately 10 km (Lafleur 1986; Gupta and Grant 1985). The batholith consists of 2 main phases: 1) an older, foliated to gneissic, hornblende-biotite tonalite-granodiorite phase; and 2) a younger weakly to unfoliated granodiorite-tonalite phase with quartz monzodiorite affinities (Lafleur 1986). The LREE are enriched 20 to 50 times chondritic values. The HREE are enriched 1 to 3 times chondritic values (Lafleur 1986).

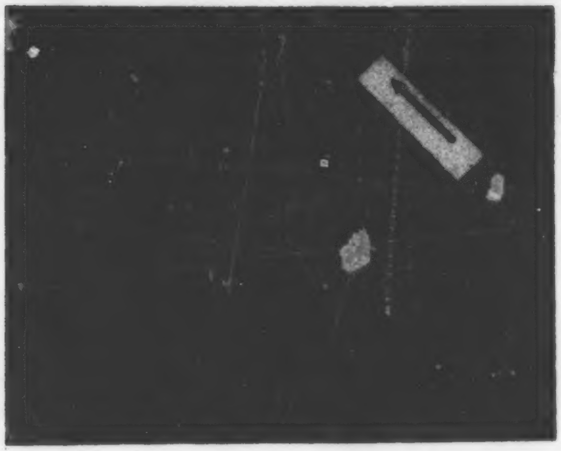


Figure 11.31. Highly strained tonalite from high-strain margin of the Round Lake batholith, Pacaud Township.

A weakly foliated part of the older phase of the batholith is ~2703 Ma (Mortensen, in prep.); a granodiorite of a younger phase is ~2697 Ma (Mortensen, in prep.). The final emplacement of the Round Lake batholith postdates the ~2701 million-year-old (Corfu et al. 1989) Skead assemblage which wraps around the batholith. The Round Lake batholith predates unfoliated granodiorite dikes which cut the highly strained margin (Figure 11.31) of the batholith and that are texturally indistinguishable from the 2697 million-year-old granodiorite (Jackson and Harrap 1989). Some boulders within the Timiskaming assemblage near Kirkland Lake resemble the gneissic phase of the Round Lake batholith, suggesting that the batholith was unroofed during deposition of the Timiskaming assemblage.

Most tonalitic phases have a foliation defined by biotite and lenticular feldspar and quartz. Near the margin of the batholith, the foliation is well developed, mylonitic and subparallel to the batholith margin (*see* Figure 11.31). Within the batholith, the foliation is variably oriented (e.g., north striking to the south of the Otto Stock and east striking near the eastern margin of the batholith).

The lack of a significant thermal aureole, the presence of steeply plunging stretching lineations and asymmetric folds of schistosity within the Pacaud structural complex that "climb" away from the batholith, and the layer-parallel mylonite zone within the batholith, all suggest that the Round Lake batholith was emplaced at its final level as a solid-state diapir (Lafleur 1986; Jackson and Harrap 1989). It is important to note that the asymmetry of minor folds of schistosity within the strain margin of the batholith is inconsistent with a cross folding origin for the elliptical map pattern of the Round Lake batholith. The internal foliation pattern of the Round Lake batholith, however, is not simply concentric as might be expected from a simple first order diapir (e.g., Schwerdtner et al. 1978). The complex foliation pattern could be explained by second-order diapiric structures budding off the main diapir, with the foliation having been developed synchronously with diapirism. Alternatively, diapirism, with second-order diapirs, of a previously subhorizontally foliated mass could explain the pattern (e.g., Schwerdtner et al. 1978; *also see* patterns displayed by salt diapirs of central Iran *in* Talbot and Jackson 1987). The details of the structural evolution of the margin of the batholith and the adjacent highly deformed supracrustal rocks are not known. The high-strain aureole may have a complex history. The possibility of the supracrustal rocks having been structurally emplaced over the Round Lake batholith before diapirism cannot be ruled out (e.g., van Staal and Williams 1983).

Lake Abitibi Batholith

The Lake Abitibi Batholith (*see* Figure 11.2) is roughly 35 by 50 km and includes hornblende diorite to quartz diorite, hornblende-biotite tonalite and granodiorite, granodiorite and tonalite, and megacrystic granodiorite (~2689 Ma, Mortensen 1987a) phases (Smith and Sutcliffe 1988). A gabbro to gabbronorite intrusion was emplaced into the hornblende-biotite tonalite and granodiorite phase of the Lake Abitibi Batholith (Smith and Sutcliffe 1988).

The REE patterns of Lake Abitibi Batholith granitoid rocks are LREE enriched (~10 to 50 times chondrite) relative to the HREE (~1 to 6 times chondrite) and indicate no significant europium anomaly (Sutcliffe et al. 1990). Hornblende from the batholith is more fractionated than rocks of the batholith (Sutcliffe et al. 1990). Lake Abitibi Batholith rocks also have high chromium and nickel contents (greater than 100 ppm). Sutcliffe et al. (1990) used these and other chemical characteristics of similar intrusions to conclude that the Lake Abitibi Batholith was derived from LIL-enriched mantle and evolved, in part, by hornblende fractionation.

Kenogamissi Batholith

The Kenogamissi Batholith (*see* Figure 11.2) separates the Abitibi greenstone belt from the Swayze area and consists of: 1) hornblende- and/or biotite-bearing granodiorite to tonalite gneiss; and 2) foliated to massive hornblende- and/or biotite-bearing granodioritic, dioritic and granitic felsic plutonic rocks (Pyke 1978; Card 1979; Bright 1984). Other than the general distribution of units, little is known about the batholith. Applying the relationships established in the comparable Round Lake and Lake Abitibi batholiths, granodiorite to tonalite gneiss is presumed to be older than hornblende-granodiorite to diorite. The batholith was emplaced into at least the Peterlong and the Eldorado (~2727 Ma) assemblages, which appear to envelope this batholith.

LATE ALKALIC PLUTONS

Alkalic plutons in the Ontario segment of the western Abitibi Subprovince occur mainly in the southern Abitibi greenstone belt; however, some syenitic rocks intrude the Pontiac Subprovince in Quebec (e.g., MERQ–OGS 1983). Some of these plutons, for example the Lebel Stock (Abraham 1951), were clearly emplaced prior to the development of the Larder–Cadillac shear zone. The plutons consist chiefly of syenite to quartz syenite, but show considerable variation. For example, the 2680±1 million-year-old (Corfu et al. 1989) Otto Stock consists chiefly of hypersolvus syenite and porphyritic syenite and minor quartz syenite, clinopyroxene-rich biotite-bearing hornblendite to diorite (Lawton 1959; Smith and Sutcliffe 1988). The syenite plutons are, in general, characterized by a low mafic mineral content (less than 9%), albite and microcline in perthitic intergrowths, euhedral feldspars and Na_2O greater than K_2O (Lawton 1959). Nepheline, indicative of silica undersaturation, is confined to the contact zone "injection gneiss" (Lawton 1959) at the northern margin of the Otto Stock. REE patterns are typically LREE enriched (20 to 500 times chondrite) relative to HREE (1 to 20 times chondrite) with the mafic rocks displaying more highly fractionated patterns than the granitoid rocks (Sutcliffe et al. 1990). Late stage oxidizing magmatic fluids are indicated by magnesium enrichment of clinopyroxene, the paucity of ilmenite and the preponderance of magnetite and primary titanite (*see* Rowins et al., in press, study of the Murdock intrusion). Sutcliffe et al. (1990) concluded that these syenitic intrusions are likely derived by melting of LIL-enriched mantle.

PORPHYRY BODIES

Numerous small (outcrop-size to several kilometres in dimension) porphyry bodies are present within the southern Abitibi greenstone belt. They vary in shape and style from dikes and sills to irregular pods (e.g., Thomson 1950). The most commonly reported types of porphyry are: 1) quartz porphyry; 2) feldspar porphyry; 3) "crowded" feldspar porphyry; 4) quartz-feldspar porphyry; and 5) augite-bearing porphyry. The groundmass to the porphyritic phases is commonly aphanitic and pink.

The porphyry bodies intrude most rock types in the southern Abitibi greenstone belt. Some porphyry dikes are deformed in the Round Lake batholith strain aureole (Jackson and Harrap 1989), dated at ~2700 Ma, and some crosscut the strain aureole. In the Timmins area, many of the larger porphyry bodies have been dated at ~2690 Ma (Corfu et al. 1989), ~10 Ma older than the Timiskaming-type sedimentary rocks in the area (Corfu et al., in press; *see* Three Nations Assemblage). In the Kirkland Lake area, Thomson (1950) reported porphyry bodies that intrude the ~2677 million-year-old Timiskaming assemblage metavolcanic and metasedimentary rocks. Thus, the duration of emplacement of the porphyry bodies spans at least 20 my from 2700 to 2680 Ma.

Mafic-Ultramafic Intrusions

A number of mafic-ultramafic intrusions occur within the Abitibi greenstone belt. These intrusions are, in general, not well documented. Some of these are described here to present a general indication of their characteristics. In the Wabigoon Subprovince, a subdivision of ultramafic to mafic intrusions has been erected (*see* Blackburn et al., this volume). Sufficient data are not available to test the application of their subdivisions in the Abitibi Subprovince. However, ultramafic-mafic intrusions occur in association with many rock types in virtually all assemblages. For example, there are some associated with mafic-ultramafic assemblages (e.g., Stoughton–Roquemaure assemblage) whereas some others are associated with intermediate to felsic metavolcanic assemblages (e.g., Blake River assemblage; *see* MERQ–OGS (1983) for distribution). In addition, there are also some gabbros associated with large felsic intrusions (e.g., Lake Abitibi Batholith; Smith and Sutcliffe 1988).

MUSKASENDA GABBRO

The Muskasenda gabbro intrudes along the contact between the Peterlong and Bartlett assemblages (*see* Figure 11.7). The Muskasenda gabbro is a north-striking, layered mafic intrusion approximately 3 by 15 km in dimension (Pyke 1978; Bright 1984). The intrusion is of tholeiitic affinity (Pyke 1978) and consists of gabbro, melanocratic gabbro, gabbroic anorthosite and locally glomeroporphyritic gabbro.

LINCOLN–NIPISSING PERIDOTITE-GABBRO INTRUSION

The Lincoln–Nipissing peridotite is a 20 km long, 1 km wide, steeply dipping intrusion that extends in a southeasterly direction from the Lebel Stock (Abraham 1951; Hewitt 1951). The peridotite is serpentinized, but commonly retains the euhedral outlines of original olivine grains (*see* Abraham 1951, Photo on p.23). The intrusion occurs near the boundary between the ~2701 million-year-old (Corfu et al. 1989) Skead assemblage and ~2705 million-year-old (Corfu et al. 1989) Larder Lake assemblage. The intrusion is also cut by both the Boston fault and the Lincoln–Nipissing shear zone. The Boston fault predates the ~2680 million-year-old Lincoln–Nipissing shear zone, thus, the Lincoln–Nipissing peridotite was emplaced sometime between 2680 and 2700 Ma, prior to some of the regional deformation.

BLACK–GRENFELL GABBRO SUITE

Mafic intrusions, including diorite, gabbro, quartz-diorite and quartz gabbro, are common east of the Watabeag Batholith (Thomson 1950; Lovell 1971). Little is known of these intrusions, but they may be related to mafic metavolcanic rocks of the Kinojevis South assemblage. The Kinojevis North assemblage does not display such widespread intrusions.

KAMISKOTIA GABBROIC COMPLEX

The Kamiskotia Gabbroic Complex occurs within the Kamiskotia assemblage. The intrusion consists of gabbroic, anorthositic and granophyric rocks; these are described in the section entitled Kamiskotia Assemblage.

MONTCALM GABBROIC COMPLEX

The Montcalm Gabbroic Complex cuts volcano-sedimentary rocks in the Belford–Strachan area of the western Abitibi greenstone belt. A more extensive description of this 2702 million-year-old (Barrie and Naldrett 1990) stratiform, layered intrusion is given in the section entitled

Belford–Strachan Area. The intrusion consists of pyroxenite, gabbro, anorthositic gabbro and ferroan gabbro zones (Barrie and Naldrett 1990). The age of the intrusion suggests that it may be coeval with spatially associated mafic and felsic metavolcanic rocks.

GHOST RANGE COMPLEX

The Ghost Range Complex lies just north of the Porcupine–Destor deformation zone (*see* Jensen 1982; Jensen and Langford 1985). The Ghost Range Complex has been variably interpreted as a unit of flat-lying lava flows (Jensen and Langford 1985) or a differentiated sill (Satterly 1952b, 1954; MacRae 1969). Recent mapping supports an origin as a layered intrusion preserved in a subhorizontal upright position (Smith and Sutcliffe 1988). The The Ghost Range Complex consists of shallow-dipping (less than 50°) units of lherzolite (with dunite), clinopyroxenite, gabbronorite (and gabbro) and hornblende-gabbro (Satterly 1952b, 1954; MacRae 1969; Smith and Sutcliffe 1988). Within the complex, a thin unit of layered felsic tuff is reported to occur between 2 gabbroic-textured phases (Jensen and Langford 1985). The origin of this felsic unit is critical to the understanding of the Ghost Range Complex, since a volcanic origin would support a flow interpretation for the complex (e.g., Jensen and Langford 1985) as opposed to an intrusive origin (e.g., Smith and Sutcliffe 1988). The Ghost Range layered intrusion may be closely related genetically to other layered, tholeiitic intrusions in this area, including the Centre Hill Complex (MacRae 1969), which occur within the Kidd–Munro assemblage (*see* Kidd–Munro Assemblage).

The Ghost Range Complex lies upon calc-alkalic basalts of the Duff–Coulson–Rand assemblage, although the contact relationship with these metavolcanic rocks is poorly constrained (Smith and Sutcliffe 1988). While shallow-dipping units do occur elsewhere in this eastern part of the Abitibi greenstone belt, the anomalously shallow dips of the Ghost Range Complex and its inferred, disrupted contact with the underlying, more steeply dipping calc-alkalic metavolcanic rocks (Satterly 1952b, 1954), suggest that the present geometry of the Ghost Range intrusion may be partly a result of structural emplacement on top of the calc-alkalic metavolcanic rocks.

MUNRO SILL (CENTRE HILL COMPLEX)

The Munro sill is a thickly(?) layered intrusion exposed in 3 principal areas, each with similar rock sections (Satterly 1953; MacRae 1969): 1) Centre Hill; 2) McCool Hill; and 3) a broad area at the boundary between Warden and Munro townships. Aeromagnetic maps indicate the Munro sill is a continuous body. The sill is approximately 11 km in length and 0.5 to 1 km thick. It is folded along a major east-trending syncline having almost vertical limbs, and is complexly disrupted by both longitudinal and cross faults. Minor folds are common on both limbs, especially between Centre Hill and the Warden–Munro areas.

The Centre Hill outcrop area reveals a complete, vertically dipping section of the Munro sill. The intrusion generally strikes easterly, but it is folded sharply southward at its western end where MacRae (1969) interpreted it to be dragged against, and cut off by, the Centre fault. Along the northern upper contact, the intrusion cuts a mafic fragmental rock (MacRae 1969). To the south, the sill lies in contact with basalt flows. The contacts are sharp and local intertonguing of the intrusive and volcanic rocks was reported (MacRae 1969). Immediately adjacent to the lower (or southern) contact, the intrusive rock may have been slightly chilled, but an exact interpretation is difficult because of the subsequent metamorphism and alteration (e.g., MacRae 1969).

The Centre Hill Complex consists of 7 cyclic layers of ultramafic and mafic rocks (MacRae 1969). The lower 5 cycles consist of successive layers of peridotite and clinopyroxenite, the sixth unit is composed of clinopyroxenite and gabbro and the upper most seventh unit consists of a layer of melanocratic gabbro overlain by normal gabbro (MacRae 1969). The ultramafic layers range in thickness from 1 to 50 m and lie in sharp contact with each other. The uppermost gabbroic unit is over 200 m thick. At the lower contact, there is a unit of hornblendite 10 to 30 m thick (MacRae 1969).

The uppermost gabbro unit is overlain by a fragmental rock described as a pillow breccia, hyaloclastite, or scoria, which is up to 75 m in thickness in the east, but becomes progressively thinner along strike to the west (Coad 1976). The eastern extension of the mafic fragmental rock is not exposed because of extensive drift cover and major north-trending faults along the eastern edge of the Centre Hill Complex (Coad 1976). This fragmental rock unit was mapped as a rhyolite "agglomerate" (Satterly 1952a); however, it has a composition intermediate between a tholeiitic olivine basalt and picritic basalt (Coad 1976). Coad (1976) argued that the contact between the pillow breccia and gabbro is faulted; however, some localized segments of the contact appear to be primary and suggest a coeval relationship exists between the intrusion and the adjacent mafic fragmental rock (e.g., Green and MacEachern 1990).

Lamprophyre Dikes

Lamprophyre dikes are commonly reported from the southern Abitibi greenstone belt (*see* Thomson 1943, 1946, 1948, 1949, 1950; Hewitt 1951; McNeil and Kerrich 1986; Wyman and Kerrich 1987, 1988; Barrie 1990b). The most common lamprophyre dikes are biotite-phenocrystic and mafic in composition. Chlorite is common in the matrix. Some dikes have chloritic clots that may represent altered mafic metavolcanic inclusions. Many of the dikes have inclusions of tonalite and/or syenite that are well rounded. The lamprophyre dikes commonly contain carbonate; it is unclear if the carbonate in these dikes is a reflection of the original volatile content or if the carbonate reflects a secondary process.

In the Kamiskotia area, an ultramafic lamprophyre suite dated at ~2687 Ma is spatially related to the Porcupine–Destor deformation zone (Barrie and Davis 1990). This suite contains biotite-lamprophyre dikes, diopside-rich

Table 11.1. Generalized assemblage types recognized in the western part of the Abitibi Subprovince, based chiefly on the presence of predominant rock types and rock associations.

A) Komatiite- and/or Tholeiite-bearing Assemblages

1. dominated by komatiite-tholeiite with interflow iron formation (e.g., Stoughton–Roquemaure)
2. dominated by komatiite and/or tholeiite with significant felsic metavolcanic rocks (e.g., Kidd–Munro)
3. dominated by komatiite-tholeiite without significant iron formation or significant felsic metavolcanic rocks (e.g., Larder Lake)
4. dominated by alternating magnesium- and iron-rich tholeiitic units (e.g., Kinojevis North and Kinojevis South)
5. dominated by thick magnesium- and/or iron-rich tholeiitic units ± minor komatiitic units (e.g., Catharine–Pacaud)
6. dominated by komatiite and/or tholeiite with felsic metavolcanic rocks and iron formation (e.g., Boston)

B) Intermediate to Felsic Metavolcanic–dominated Assemblages

1. dominated by intermediate to felsic fragmental units (e.g., Skead)
2. dominated by intermediate effusive metavolcanic units (e.g., Deloro)

C) Clastic Metasedimentary–dominated Assemblages

1. dominated by turbiditic metasedimentary units (e.g., Hearst)
2. dominated by alluvial-fluvial metasedimentary ± alkalic metavolcanic units (e.g., Timiskaming)

dikes, garnet-rich dikes and pseudoleucite-bearing dikes. The suite is characterized by extreme LIL and LREE enrichment (Barrie 1990b). An altered ~2673 million-year-old (Corfu et al. 1989) albitite dike present in the Hollinger Mine stratigraphy (Tisdale assemblage) has lamprophyre affinities (Wood et al. 1986).

DISCUSSION

Assemblages

In the southern Abitibi greenstone belt, excluding the 2685 to 2675 million-year-old Timiskaming-like assemblages, supracrustal assemblages range in age from ~2750 to 2700 Ma. Most assemblages are younger than 2720 Ma. The older assemblages (2720 to 2750 Ma) tend to be exposed adjacent to large batholithic complexes, such as the Bartlett and Catharine–Pacaud assemblages in Ontario and the Hunter Mine assemblage in Quebec. Based chiefly on the presence of major rock types and rock associations, the following recognized general assemblage types are summarized in Table 11.1.

At a large scale, the southern Abitibi greenstone belt can be divided into 2 domains separated roughly by the trace of the Porcupine–Destor deformation zone. South of the shear zone, the belt is underlain principally by tholeiite and calc-alkalic intermediate to felsic assemblages and subordinate komatiite-bearing assemblages. North of the Porcupine–Destor deformation zone, the southern Abitibi greenstone belt is underlain chiefly by metavolcanic rocks of komatiitic to tholeiitic chemical affinity (including intermediate to felsic rocks).

SUMMARY OF ROCK ASSEMBLAGES SOUTH OF THE PORCUPINE–DESTOR DEFORMATION ZONE

South of the Porcupine–Destor deformation zone, many dated metavolcanic assemblages are younger than ~2705 Ma. Exceptions include the Catharine–Pacaud and Bartlett assemblages, which predate 2720 Ma, and possibly the Boston, Peterlong, Eldorado and Deloro assemblages. Assemblages from the Shining Tree and Swayze areas, other than the Ridout assemblage, are not discussed here, because insufficient geological and geochronological data exist to make valid correlations or generalizations.

The older (pre–2720 Ma) and presumed older assemblages can be grouped into: 1) those containing mainly tholeiitic metavolcanic rocks not displaying decametre-scale layering of iron- and magnesium-rich tholeiitic basal units (e.g., Peterlong and Catharine–Pacaud assemblages); and 2) those containing iron formation and felsic calc-alkalic metavolcanic rocks with variable amounts of ultramafic and tholeiitic mafic metavolcanic rocks (e.g., Eldorado, Boston and Bartlett assemblages).

Although the Bartlett and Eldorado assemblages may be correlative, their presumed substrata are not. The Peterlong assemblage (substratum to the Bartlett assemblage) is a tholeiitic unit and the Deloro assemblage (substratum to the Eldorado assemblage) is a calc-alkalic assemblage. Thus, if the Bartlett and Eldorado assemblages are correlative, then either the base of the Eldorado and Bartlett assemblages is a major tectonic contact, or it on-laps different basal assemblages.

The younger, ~2705 to 2700 Ma, and presumed younger assemblages form 4 distinct groups. 1) The Bowman and Larder Lake assemblages are chiefly komatiitic and tholeiitic metavolcanic rocks and are ~2705 Ma. 2) The Kinojevis North, Kinojevis South and Geikie assemblages consist of magnesium- and iron-rich units, interbedded on a scale of 10s of metres, belonging to a tholeiitic basalt suite. In contrast, the older tholeiitic assemblages (Catharine–Pacaud and Peterlong) have magnesium- or iron-rich formations several kilometres in apparent thickness. 3) Several calc-alkalic felsic fragmental assemblages (e.g., Skead, Gauthier and possibly Halliday assemblages) are ~2700 Ma. These assemblages contain neither pillowed intermediate members nor iron formation, whereas the older felsic assemblages do contain iron formation. The felsic fragmental members of the Blake River and Watabeag assemblages in Ontario may be grouped with the Skead, Gauthier and Halliday assemblages because of their calc-alkalic character. However, the Watabeag and the Blake River assemblages differ from these assemblages in that they contain substantial intermediate to mafic pillowed members. Note also that in Quebec, tholeiitic rocks are included in the Blake River assemblage (e.g., MERQ–OGS 1983). 4) The youngest rock assemblages include the Timiskaming, Midlothian and Ridout assemblages. These assemblages are characterized by alluvial-fluvial facies to resedimented turbiditic facies. In the Kirkland Lake–Larder Lake area, these assemblages contain alkalic metavolcanic rocks. They unconformably overly previously deformed pre–2700 Ma metavolcanic units.

SUMMARY OF ASSEMBLAGES NORTH OF THE PORCUPINE–DESTOR DEFORMATION ZONE

North of the Porcupine–Destor deformation zone in Ontario, assemblages range in age from approximately less than 2730 to 2700 Ma. The Stoughton–Roquemaure assemblage is assumed to be the oldest assemblage as it apparently overlies the *circa* 2730 million-year-old (Mortensen 1987a) Hunter Mine group exposed in Quebec. The Kidd–Munro assemblage is *circa* 2717 to 2715 Ma (Barrie and Davis 1990; Corfu et al. 1989). The Stoughton–Roquemaure and Kidd–Munro assemblages are dominated by komatiitic and tholeiitic metavolcanic rocks and coeval intrusions. The Kidd–Munro assemblage, however, contains units of oxide-facies iron formation or sulphidic interflow metasedimentary rocks and units of icelandite and high-silica rhyolite (Type FIII, Lesher et al. 1986). Although minor units of silica-rich rhyolite do occur in the Stoughton–Roquemaure assemblage (Jensen and Langford 1985), available geochemical data are limited and prohibit their classification as high-silica rhyolite (type FIII), as defined by Lesher et al. (1986).

The 2707 to 2705 million-year-old Kamiskotia assemblage (Barrie and Davis 1990) and the 2700 million-year- old top of the Krist assemblage (Corfu et al. 1989) are the youngest metavolcanic assemblages north of the Porcupine–Destor deformation zone. Icelandite and high-silica rhyolite also occur in the younger Kamiskotia assemblage, but no komatiitic metavolcanic rocks occur. The Tisdale assemblage, interpreted to be slightly older than the 2700 million-year-old Krist assemblage, contains komatiitic and tholeiitic rocks; however, mafic and ultramafic intrusions, icelandite, high-silica rhyolite and oxide-facies iron formation are absent from the Tisdale assemblage. The 2700 million-year-old Krist assemblage consists of intermediate to felsic fragmental calc-alkalic metavolcanic rock, similar to the Skead and Gauthier assemblages.

STRATIGRAPHIC IMPLICATIONS

The majority of the assemblages discussed above are either in fault contact with adjacent assemblages or the contact relationship is unknown (*see* Figure 11.8). The obvious exception is the unconformable relationship between the Timiskaming and Timiskaming-like assemblages and adjacent pre–2700 Ma assemblages. Even where depositional contacts seem probable, they have been modified by tectonism. In most cases, the nature of the fault separating adjacent assemblages is unknown and the adjacent assemblages are quite distinct, even though similar rock types may be found in each. These conclusions make it difficult to establish rigorous regional stratigraphic models.

Older assemblages occur near the margins of large batholithic complexes (Round Lake and Kenogamissi batholiths). These older assemblages may have been "dragged" to higher stratigraphic and structural levels during the emplacement of these intrusions. The distribution of these assemblages from Timmins to south of the Round Lake batholith suggests a widespread, somewhat older (pre–2720 Ma) metavolcanic substratum to the post–2705 Ma assemblages. The presence of old (~2755 Ma; Corfu et al., in press) xenocrystic zircons in a lamprophyre dike, which cuts the Timiskaming assemblage, may support this interpretation or alternatively indicate a different older source.

The younger metavolcanic assemblages in the southern Abitibi greenstone belt may be underlain by older metavolcanic assemblages; however, the regional nature of the contact is not well studied. In the Shaw Dome area, a komatiite of the Bowman assemblage lies stratigraphically upon iron formation of the Eldorado assemblage (~2730 Ma inferred) (Pyke 1982). The contact between these 2 assemblages is, however, locally marked by a thrust fault (Piroshco and Kettles 1987, 1988, 1991) and the Eldorado assemblage is locally highly folded and possesses well-developed schistosity (Pyke 1982). East of the Kenogamissi Batholith, the boundary between the Bartlett assemblage (~2730 Ma) and younger Geikie assemblage (~2700 Ma) appears to be a structural discordance and is locally marked by a well developed shear zone (R.B. Hrabi, Queen's University, personal communication, 1990). At the same boundary, a conglomeratic unit is present (R. Shegelski, Esso Minerals, personal communication, 1989; R.B. Hrabi, Queen's University, personal communication, 1990) which may represent a basal conglomerate to the Geikie assemblage. Finally, east of the Round Lake batholith, the boundary between the Skead (2700 Ma) and Catharine–Pacaud (2720 Ma) assemblages is interpreted as a sheared primary contact (Jackson and Harrap 1989). In summary, the contact between the younger (post–2705 Ma) and older (pre–2720 Ma) metavolcanic assemblages appears to be a tectonically reactivated primary contact; however, more mapping is required to substantiate this.

Structural Geology

Tectonic fabric and mesoscopic structures are domainal in distribution and type within the southern Abitibi greenstone belt. In general, fabric and structures parallel, and are best developed adjacent to regional faults, large batholiths and many of the exposed assemblage boundaries. Limited data preclude a rigorous belt-wide synthesis. Although there is considerable detail available for some areas (e.g., Hamilton 1986; Piroshco and Kettles 1991), it is unclear whether the results of these detailed studies can be applied throughout the greenstone belt. Nevertheless, some success has been made in regional structural analysis (e.g., Hodgson and Hamilton 1989).

SUBPROVINCE BOUNDARIES

The structural style of the boundaries of the Abitibi Subprovince in Ontario, with the exclusion of the boundary with the Wawa Subprovince (*see* Williams et al., this volume and references therein) are not well understood. The boundary with the Pontiac Subprovince is present only in Quebec (*see* Dimroth et al. 1982, 1983a, 1983b). The boundary with the Ramsey–Algoma granitoid complex is not well documented, but may be intrusive. Finally, the boundary with the Opatica Subprovince in Ontario is not exposed.

CHRONOLOGY OF STRUCTURAL EVENTS IN THE SOUTHERN ABITIBI GREENSTONE BELT

In general, structures within the southern Abitibi greenstone belt can be assigned to one of the following: a) "pre-cleavage" folds (these are the earliest recognized structures); b) structures related to batholith emplacement at ~2700 to 2690 Ma; and c) syn– to post–Timiskaming structures.

Only where crosscutting or absolute-timing constraints exist can any structure be placed into the above scheme. Many structures cannot be precisely placed in this scheme. It is emphasized that the classification of structures is incomplete; for example, there could be a class of structures that formed between "b" and "c" above; but which, at present, has not been documented.

"Pre-cleavage" folds, some thrust faults, emplacement of large granitoid complexes and formation of related structures predated deposition of the 2685 to 2675 million-year-old Timiskaming-like assemblages. Synchronous with, to postdating deposition of, the Timiskaming-like assemblages, prominent regional shear zones and fold sets developed, including a northwest- to west-northwest-striking set that, locally, predates a northeast- to east-northeast-striking set (e.g., Hodgson et al. 1990). Thrust faulting and/or steep reverse faulting also occurred at this time. These structures can be interpreted to have formed during late-Archean, protracted north-south subhorizontal compression (Dimroth et al. 1983b; Hodgson et al. 1990).

The above structures are overprinted by north-east-, northwest- and north-northeast-striking brittle faults related to the development of the Paleoproterozoic Cobalt Embayment and the Phanerozoic Timiskaming Rift.

EARLY FOLDING, THRUSTING AND EMPLACEMENT OF LARGE GRANITOID MASSES

In the Timmins area, the Porcupine Syncline is unconformably truncated by the base of Timiskaming-like metasedimentary rocks (Three Nations assemblage). No cleavage is associated with this fold, and the earliest regionally recognized cleavage in the area transects the axial trace of the fold (Hodgson 1982; Roberts 1981). Pre–Timiskaming folding is also well documented in the Kirkland Lake–Larder Lake area, where pre–2700 Ma volcanic rocks were folded and then unconformably overlain by the Timiskaming assemblage (Thomson 1950; Hewitt 1963). The orientation, significance and distribution of these early folds is not known and will be difficult to completely establish since they are overprinted by all the later structures.

Thrust faulting occurred early in the structural history of the greenstone belt. In the Timmins area, Piroshco and Kettles (1988, 1991) documented a shallowly north-dipping shear zone with an interpreted reverse sense of shear. The thrust fault occurs at the boundary between the Deloro and Eldorado assemblages, south of the Porcupine–Destor deformation zone, and may extend around the Shaw Dome (*see* Deloro Assemblage). Other thrust faults are interpreted to occur within the Eldorado assemblage above the thrust separating the Deloro and Eldorado assemblages (Piroshco and Kettles 1987, 1988, 1991).

East of the Round Lake batholith, a fault at the base of the Catharine formation within the Catharine–Pacaud assemblage has a geometry consistent with, but not proof of, a south-verging flat-ramp-flat thrust geometry (Jackson and Harrap 1989). This fault predated emplacement of the Round Lake batholith at ~2700 Ma.

The Round Lake batholith was emplaced after formation of the ~2701 million-year-old (Corfu et al. 1989) Skead assemblage, which is deflected by the batholith, and before emplacement of a 2697 million-year-old (Mortensen, in prep.) granodiorite that is mineralogically and texturally identical to dikes that cut the high-strain margin of the batholith. Emplacement of the batholith induced strong deformation in supracrustal rocks within 1 to 2 km of the batholith (Jackson and Harrap 1989). Outside the high-strain aureole, supracrustal rocks are not penetratively deformed; however, emplacement of the batholith warped and folded adjacent and outlying strata on a regional scale. For example, the ~2700 million-year-old Skead assemblage can be traced in a semicontinuous arc around the eastern margin of the batholith for over 70 km. Throughout this arc, the strata are steeply dipping to overturned (e.g., Johns 1982). Structures related to batholith emplacement are younger than the 2700 million-year-old Skead assemblage, but older than the inferred ~2697 Ma age of the granodiorite dikes. Consequently, the structures predate the 2685 to 2765 million-year-old Timiskaming assemblage and subsequent deformation by at least ~15 to 25 my.

The early diapiric emplacement of large batholithic masses into the supracrustal rocks may account for some of the pre–Timiskaming folds that lack axial planar fabric (e.g., DA1 folds of Powell et al. 1990; and Porcupine Syncline) in the southern Abitibi greenstone belt. Note, however, that an ~15 to 25 my window between diapirism and development of the Timiskaming assemblage allows for the possibility that some of these pre–Timiskaming folds may also have formed independently of early diapirism by as yet unrecognized tectonic processes.

Some granitoid masses such as the Round Lake batholith could represent structural basement to some Abitibi greenstone belt supracrustal assemblages. Early thrusting of dense mafic metavolcanic-dominated assemblages over a more buoyant granitoid domain could have provided the necessary density inversion for diapirism. An age for the Round Lake batholith that is younger than the surrounding supracrustal assemblages does not eliminate the diapir hypothesis, since this age relationship is also consistent with older volcanic rocks having been thrust over younger structural basement. An older age for the Round Lake batholith would, however, strongly support the hypothesis that the Round Lake batholith represents diapirically emplaced structural basement.

It is not clear if any deformation predates diapirism; however, some relationships are at least consistent with this hypothesis. The internal foliation pattern of the Round Lake batholith is consistent with diapirism of a previously subhorizontally foliated mass and is comparable to patterns displayed by evaporite beds in salt diapirs of central Iran (Talbot and Jackson 1987; Jackson, Kimmerly et al. 1990). The possible fault at the base of the Catharine formation in the Catharine–Pacaud assemblage may also be a pre-diapir structure.

REGIONAL FOLDS AND SHEAR ZONES POSTDATING THE TIMISKAMING-LIKE ASSEMBLAGES

Regional folds in the southern Abitibi greenstone belt tend to be upright, shallowly plunging to doubly plunging structures (Hodgson and Hamilton 1989; Hodgson et al. 1990). The 2 most prominent sets of folds and shear zones in the southern Abitibi greenstone belt are an early northwest- to west-northwest-striking set and a later northeast- to east-northeast striking set (e.g., Hodgson and Hamilton 1989; Hodgson et al. 1990). Deflection of AEM conductors about the northwest-trending Bradburn–Coulson and Cochrane–Milligan shear zones (*see* Figure 11.9) suggests a dextral component of horizontal slip. These shear zones either terminate at or merge with the Porcupine–Destor deformation zone south of the Lake Abitibi Batholith (*see* Figure 11.9). Jackson and Harrap (1989) also interpreted the northwest-striking Catharine shear zone to have a dextral component of horizontal slip.

The most prominent shear zones in the region are the Porcupine–Destor deformation zone and the Larder–Cadillac shear zone (*see* Figure 11.9). These shear zones transect the Abitibi greenstone belt in a roughly easterly direction for several hundred kilometres. They juxtapose markedly different supracrustal assemblages along their length, and are spatially associated with the youngest supracrustal assemblages of the greenstone belt—Timiskaming-like assemblages.

In the Timmins area, the steeply south-dipping Porcupine–Destor deformation zone juxtaposes relatively young Timiskaming-like rocks to the north against older metavolcanic rocks to the south, which is consistent with a post–Timiskaming vertical south-over-north slip (Hodgson et al. 1990). This is consistent with seismic reflection data in the Lake Abitibi area, which indicate that there is a significant component of south-side-up displacement along the south-dipping Porcupine–Destor deformation zone (e.g., Green et al. 1990; Jackson, Sutcliffe et al. 1990). Later reactivation of the Dome (Porcupine–Destor) shear zone apparently involved a component of sinistral horizontal slip, based on a left-handed rotation sense of foliation and fold axis into the Dome shear zone (Hodgson et al. 1990). Piroshco and Kettles (1991) recognized a later component of dextral, horizontal slip along the Porcupine–Destor deformation zone. Given the apparent complexity of the later deformation history along the Dome and Porcupine–Destor shear zones in the Timmins area, which was interpreted principally using deflected fabrics and fold axes, and the difficulty in correlating these fabrics across the area, the deflected fabrics and fold axes could indicate a strain regime dominated by flattening, rather than simple shear.

The Larder–Cadillac shear zone (Hamilton 1986; Toogood and Hodgson 1986; Robert 1990) is better exposed and has been more intensely studied. In Ontario, this shear zone is steeply south dipping and contains steeply south-plunging stretching lineations. The ~2705 million-year-old Larder Lake assemblage rests in the hanging wall of the shear zone and the 2685 to 2675 million-year-old generally south-facing Timiskaming assemblage forms the footwall to the shear zone. The southerly dip, steeply plunging stretching lineations and older-to-the-south on younger-to-the-north relationship all suggest that the main component of displacement was reverse and south-side-up. The Larder–Cadillac shear zone truncates the 2680 million-year-old Lebel Stock and cuts the eastern extent of the Spectacle Lakes Anticline that has Timiskaming assemblage units on both limbs. The Larder–Cadillac shear zone, then, is one of the youngest Archean structures in the greenstone belt.

LATE FAULTS RELATED TO THE COBALT EMBAYMENT AND TIMISKAMING RIFT

The Cobalt Embayment is a down-dropped crustal-scale block that infilled with sediments in Paleoproterozoic time to form the Huronian Supergroup. Faults related to basin development are north- and north-northwest-striking normal faults with an apparent sinistral component of displacement. These faults are part of a central Superior Province-wide fault set. The Timiskaming Rift (*see* Kumarapelli and Saull 1966) developed during the Phanerozoic. Some coarse clastic Proterozoic sedimentary facies are restricted to the proximity of faults related to the Timiskaming Rift, which suggests that these faults may have had a Paleoproterozoic displacement history (Rice 1988, 1991). The structural expression of the Timiskaming Rift is primarily in the form of northwest-striking normal faults; however, at the western edge of the Blake River assemblage, northwest-striking folds are present which are spatially related with a high density of northwest-striking faults. Along strike to the northwest and southeast of this region, there are regional "bends" in the Porcupine–Destor deformation zone and the Larder–Cadillac shear zone, respectively (*see* Figure 11.9). These bends may in part be related to, or accentuated by, the development of the Timiskaming Rift; although this hypothesis has not been tested. Furthermore, outliers of Timiskaming-like rocks occur in this corridor immediately south of the Porcupine–Destor deformation zone (e.g., Troop 1989) and immediately north of the main belt of the Timiskaming assemblage, northwest of Kirkland Lake. Hodgson (1983a, 1983b) noted that there was an apparent preference for Neoarchean alkalic intrusions in this region.

Metamorphism

METAMORPHIC FACIES AND METAMORPHIC CONDITIONS

Metamorphic grade of the southern Abitibi greenstone belt varies from subgreenschist facies in the area of the Blake River assemblage to greenschist facies to the north, south and west of the Blake River assemblage (Jolly 1974, 1977, 1978, 1980; Powell et al. 1990). Amphibolite facies contact aureoles are well developed near the late (~2680 Ma) alkalic intrusions and near most of the large tonalite-granodiorite bodies (e.g., Lake Abitibi Batholith; Jolly 1978, 1980). The greenschist and amphibolite facies assemblages spatially associated with granitoid intrusions apparently overprint and postdate prehnite-pumpellyite facies assemblages (Jolly 1978, 1980; Dimroth and Lichtblau 1979); although, the textural features cited could, alternatively, be explained simply by outward progradation of the thermal anomaly associated with the granitoid rocks. At least one large granitoid body, the Round Lake batholith, does not have a well-developed amphibolite facies aureole and rocks adjacent to the batholith display typical greenschist facies minerals (Jolly 1980; Jackson and Harrap 1989). Other intrusions (e.g., the Flavarian Pluton and Powell intrusion in Quebec) have mineral assemblages comparable to the adjacent greenschist facies supracrustal rocks. These intrusions are interpreted to have been emplaced prior to, or synchronous with, regional metamorphism (Goldie 1976, 1979a, 1979b).

Metamorphic facies reported by Jolly (1980) are, in part, affected by postmetamorphic faulting, since the metamorphic facies are, in part, limited by faults. The Mulven fault is such an example as prehnite-pumpellyite facies assemblages are recorded to the northwest of the fault and greenschist-prehnite-pumpellyite facies are recorded to the southeast. Powell et al. (1990) have also indicated that regional metamorphism predates at least some of the displacement along the Larder–Cadillac deformation zone.

The relationship of the Larder–Cadillac shear zone and regional metamorphic facies is not certain. At least some metamorphism in the form of large aureoles around large pre-2690 Ma granitoid bodies and smaller aureoles around smaller ~2680 million-year-old alkalic intrusions (e.g., Lebel Stock) predates the Larder–Cadillac shear zone. In addition, on a regional scale, higher metamorphic grades are found south of the Larder–Cadillac shear zone (greenschist to amphibolite) rather than north (greenschist to subgreenschist). These observations suggest that a substantial portion of movement along the Larder–Cadillac shear zone occurred during late- to post-regional metamorphism.

The physical conditions of metamorphism are not well understood for most of the Abitibi greenstone belt. The well-developed prehnite-pumpellyite facies between the 2 regional shear zones suggests a maximum temperature of approximately 300°C and metamorphic pressure on the order of 2 to 3 kbar (0.2 to 0.3 GPa; *see* P–T grid of Liou et al. 1985).

In Quebec, just north of Rouyn, Riverin and Hodgson (1980) reported cordierite-anthophyllite-bearing assemblages and andalusite-bearing assemblages in metamorphosed altered rocks associated with copper-zinc mineralization in the contact aureole of the Lac Dufault granodiorite. These observations indicate that metamorphic pressure for this region was less than ~3.5 kbar (0.35 GPa; *see* P–T grid of Carmichael et al. 1987) at the time of emplacement of the intrusion. The presence of andalusite and kyanite in a shear zone in the Bousquet Mine, Quebec (Tourigny et al. 1989), north of and parallel to the Larder–Cadillac shear zone, also indicates relatively low pressure metamorphism.

Most of the region south of the Larder–Cadillac shear zone in Ontario is of greenschist facies with epidote-actinolite-bearing assemblages being the most common. Greenschist facies assemblages indicate a minimum temperature of metamorphism of 300° to 350°C (*see* P–T grid of Liou et al. 1985). Greenschist facies and pumpellyite-epidote-bearing rocks are reported adjacent to the Round Lake batholith (Jolly 1978, 1980).

The Pontiac Subprovince, which contains 2700 million-year-old detrital zircons, was metamorphosed to much higher grade (Jolly 1978, 1980) than adjacent coeval metavolcanic rocks. The presence of staurolite, sillimanite and the migmatized nature of some of the high-grade rocks indicates that metamorphic temperature exceeded 500°C for most of the belt, but the pressure attained is unknown.

TIMING OF METAMORPHISM

The following observations bear on the timing of regional metamorphism:

1. Many of the metavolcanic rocks are ~2700 Ma. Thus, regional metamorphism postdates 2700 Ma.
2. Metamorphic grade increases towards large batholiths (e.g., Lake Abitibi batholith). Most of these bodies were emplaced between 2700 and 2690 Ma. Thus, if

regional metamorphism is associated in time and space to these intrusions, it must have predated or shortly followed 2690 Ma.

3. The Timiskaming assemblage and related rocks in Ontario and Quebec are, in part, younger than 2680 Ma (Corfu et al., in press) and were metamorphosed to greenschist facies conditions (Jolly 1978, 1980).

These observations may indicate that there were 2 metamorphic episodes, one prior to 2690 Ma associated with pre–Timiskaming deformation and emplacement of large granitoid intrusive complexes, and one after 2675 Ma associated with post–Timiskaming deformation.

Tectonics of the Western Abitibi Subprovince

Recent reviews and interpretations of the tectonics of the Superior Province indicate that crustal-scale subdivisions can be explained by plate tectonic models (e.g., Langford and Morin 1976; Hoffman 1989; Card 1990). In these models, the Shebandowan–Wawa–Abitibi greenstone belts are envisaged as the youngest, and most southerly, of a series of east-trending island arcs that were accreted to the Uchi Subprovince (Figure 11.32); metasedimentary subprovinces between the greenstone belts are generally envisaged as accretionary prisms (*see* Figure 11.32). Collision of the Minnesota River Valley gneiss terrane with the Wawa and Abitibi subprovinces is postulated to have terminated subduction beneath the Shebandowan–Wawa–Abitibi greenstone belts (e.g., Hoffman 1989).

The western Abitibi Subprovince comprises the southern Abitibi greenstone belt, the Pontiac Subprovince and the Ramsey–Algoma complex (Figure 11.33). Plate tectonic models of this region generally describe the southern Abitibi greenstone belt as an island arc and the Pontiac Subprovince as either a foreland, fore-arc or back-arc basin (e.g., Dimroth et al. 1983b; Ludden et al. 1986; Hodgson and Hamilton 1989). In these models, calc-alkalic sequences are equated with island-arc deposits and are assumed to rest on an older komatiitic to tholeiitic base (oceanic crust; Figure 11.34). The diversity of coeval supracrustal assemblages and the repetition of certain assemblages in time and space (Figures 11.35 and 11.36)in the southern Abitibi greenstone belt may indicate, however, that a more complex tectonic environment such as is now operative in the Indonesian region (Figure 11.37) was present during development of the greenstone belt.

Possible tectonic environments of formation of the supracrustal assemblages of the southern Abitibi greenstone belt are discussed in the following sections. General features of the Pontiac Subprovince and the Ramsey–Algoma granitoid complex and their relationships to each other and the southern Abitibi greenstone belt are also discussed (*see* Figure 11.33). A review of tectonic models for the western Abitibi Subprovince is given and a revised plate tectonic model is also described.

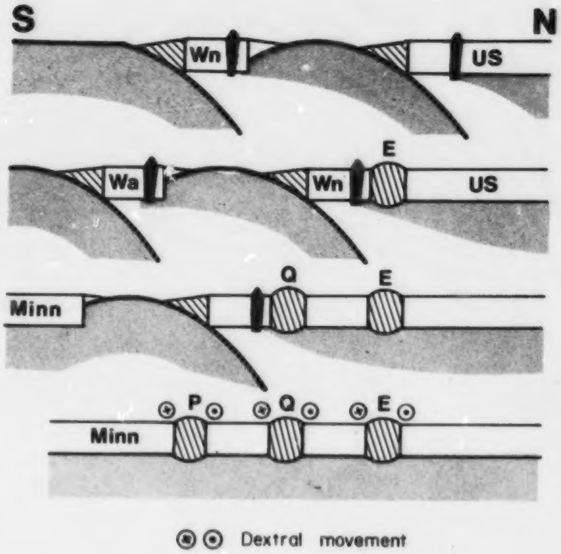


Figure 11.32. Schematic tectonic evolution of the southern Superior Province; based on Langford and Morin (1976) and Hoffman (1989). The Wabigoon (Wn) and Wawa and Abitibi subprovinces (Wa) are envisaged as island-arc complexes accreted to the Uchi Subprovince (US) involving terminal collision with the Minnesota River Valley gneiss terrane (Minn). The English River (E), Quetico (Q) and Pontiac (P) metasedimentary belts are thought to represent accretionary wedges. Late dextral transpression followed the terminal collision. Progression of events depicted from top to bottom of cartoon.

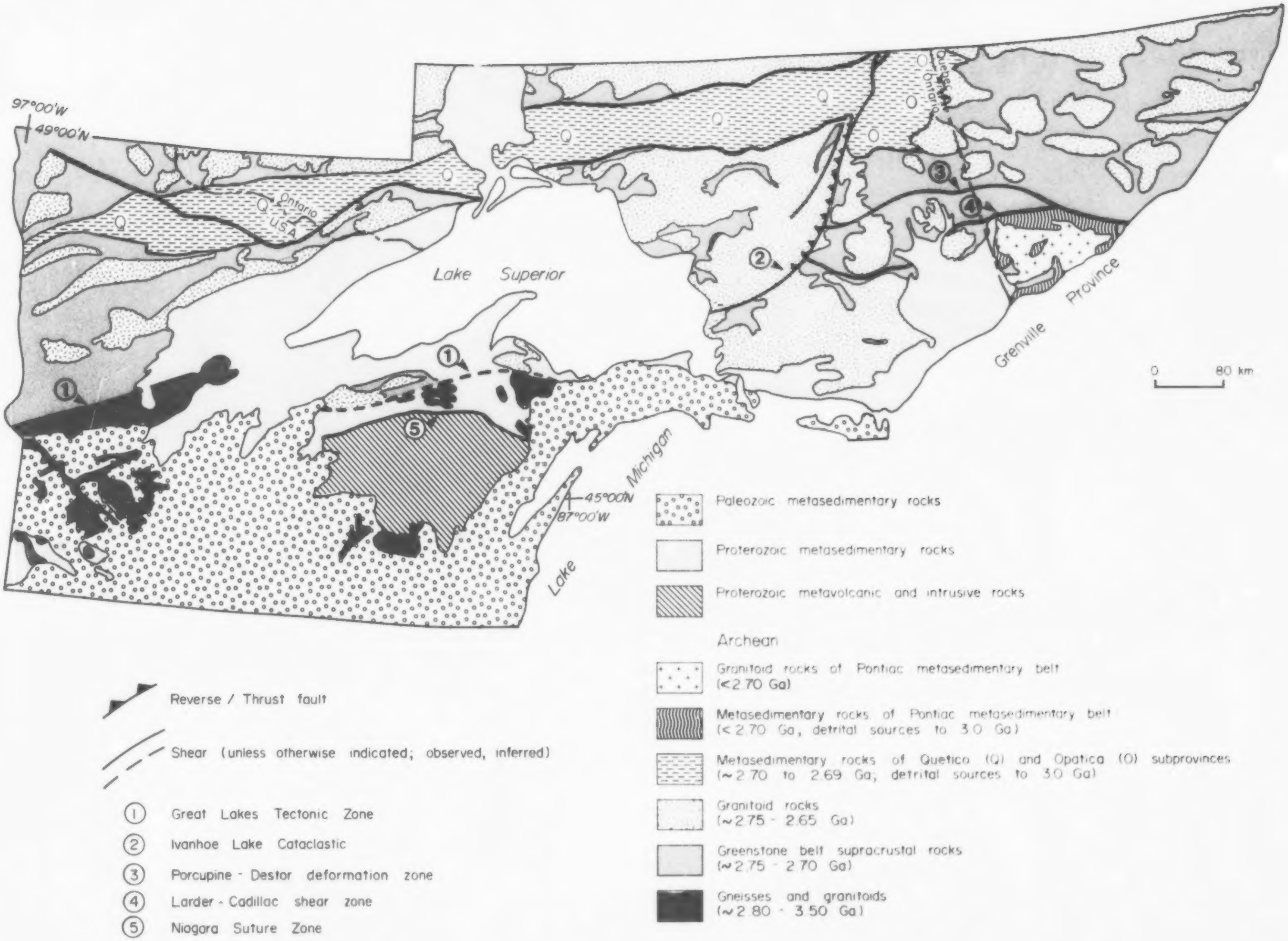
SOUTHERN ABITIBI GREENSTONE BELT

Supracrustal rocks of the southern Abitibi greenstone belt consist largely of: submarine, 2750 to 2700 Ma, ultramafic to felsic metavolcanic units; turbidites that are younger than 2700 Ma; and alluvial-fluvial metasedimentary rocks and alkalic metavolcanic rocks that are 2685 to 2675 Ma and generally confined to the proximity of regional fault zones. The environment of formation of these assemblages and their possible tectonic origin are discussed below.

2750 to 2700 Ma Submarine Ensimatic Volcanism

The supracrustal assemblages of the southern Abitibi greenstone belt have textures and primary structures that indicate formation in a submarine environment between 2750 and 2700 Ma. Few unequivocal subaerial metavolcanic rocks are documented (e.g., Comba et al. 1986; Binney and Barrie 1990) and those that have been documented as subaerial comprise a very minor component of the total area of supracrustal rock.

The role of crust and mantle in the genesis of magmatic elements of the southern Abitibi greenstone belt, for which there exist REE and/or isotopic constraints, is summarized in Table 11.2. Note that different types of data are used to constrain the interpretations. For example, lead isotope data are used for late intrusions and neodymium isotope data are used for the Kamiskotia assemblage.


97°00'W
49°00'N
Ontario
U.S.A.
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O
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Lake Superior
Michigan
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45°00'N
87°00'W
Grenville Province
0 80 km
①
②
③
④
⑤
Paleozoic metasedimentary rocks
Proterozoic metasedimentary rocks
Proterozoic metavolcanic and intrusive rocks
Archean
Granitoid rocks of Pontiac metasedimentary belt (<2.70 Ga)
Metasedimentary rocks of Pontiac metasedimentary belt (< 2.70 Ga, detrital sources to 3.0 Ga)
Metasedimentary rocks of Quetico (Q) and Opatica (O) subprovinces (~2.70 to 2.69 Ga; detrital sources to 3.0 Ga)
Granitoid rocks (~2.75 - 2.65 Ga)
Greenstone belt supracrustal rocks (~2.75 - 2.70 Ga)
Gneisses and granitoids (~2.80 - 3.50 Ga)
Reverse / Thrust fault
Shear (unless otherwise indicated; observed, inferred)
① Great Lakes Tectonic Zone
② Ivanhoe Lake Cataclastic
③ Porcupine - Destor deformation zone
④ Larder - Cadillac shear zone
⑤ Niagara Suture Zone

Figure 11.33. General geology of the southern Superior Province. Circled numbers denote major shear zones.

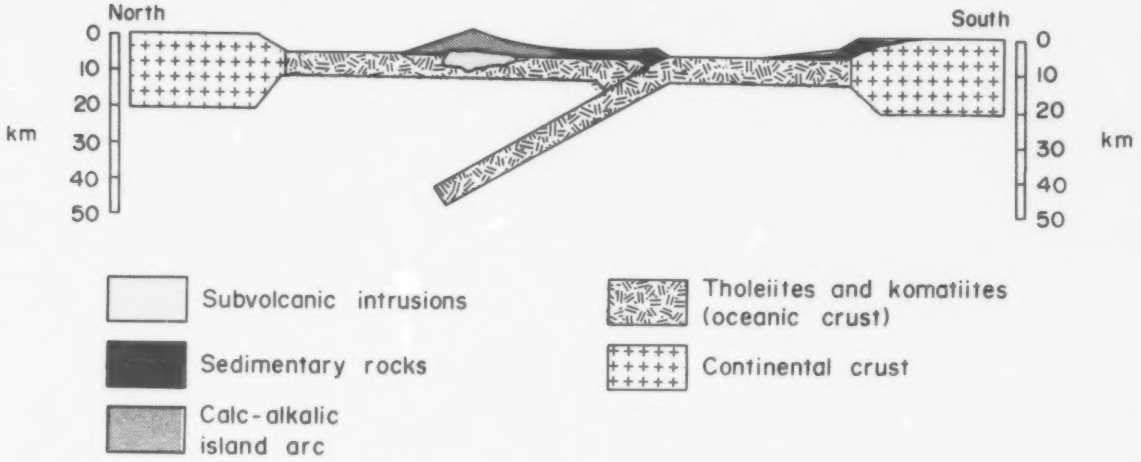

Figure 11.34. Island-arc representation of southern Abitibi greenstone belt (*after* Hodgson et al. 1990). No horizontal scale inferred.

All pre–2700 Ma assemblages can be modelled as mantle derivatives, although the inferred degree of REE enrichment and/or depletion of the mantle is variable. No older evolved crustal component is evident in their evolution (*see* Table 11.2). If there is a crustal component to these assemblages, then, the crust in question must have been extracted from the mantle just prior to formation of the southern Abitibi greenstone belt rocks. Consequently, a slightly older-arc assemblage, for example, forming basement to some of the metavolcanic assemblages, cannot be ruled out (e.g., Bowman assemblage).

The most plausible first order environment for the supracrustal assemblages of the southern Abitibi greenstone belt appears to be "ensimatic" (or "oceanic"); (e.g., Jackson and Sutcliffe 1990, and references therein). Tectonic environments suggested for the basic assemblage types summarized in the section entitled Assemblages are discussed below.

Tholeiite- and Komatiite-dominated Assemblages

In Phanerozoic oceanic tectonic settings, major accumulations of tholeiitic magma are generated at mid-ocean ridges, primitive island arcs, ocean islands and back-arc basins. An outstanding problem is whether the tholeiite- and komatiite-dominated assemblages in the Abitibi Subprovince represent different tectonic settings or different products of the same tectonic setting. For example, in an oceanic spreading environment, larger degrees of partial melting due to higher mantle temperature might generate komatiitic magma, whereas smaller degrees of partial melting due to lower mantle temperature might result in tholeiitic magma. Additionally, komatiitic magmas might be generated in a variety of tectonic settings. For example, in addition to an oceanic spreading environment, komatiitic magma can theoretically be generated by ~50% partial melting of sub-arc lithosphere hydrated by underplating of subducted oceanic lithosphere (Ghomshei et al. 1990).

Significant advances have been made in the discrimination of tectonic setting of tholeiitic rocks using trace and rare earth element contents combined with radiogenic isotope analysis. In general, such data are lacking for tholeiitic rocks of the western Abitibi Subprovince at the assemblage scale; consequently, tectonic settings can be suggested for only a limited number of assemblages. Evidence for the eruption of komatiitic flows in different tectonic regimes does, however, come from other Archean terranes. Two tectonic regimes are implied for komatiites of the Kolar Schist Belt (Rajamani et al. 1985; Krogstad et al. 1989). LREE-depleted komatiites are interpreted to have formed at Archean ocean ridges (Rajamani et al. 1985). LREE-enriched komatiites with high positive $\varepsilon_{Nd}^{(i)}$ values are inferred to have been derived from a mantle source region subjected to long-term depletion of LREE, but enriched just prior to generation of komatiite magma (Rajamani et al. 1985). This komatiite type is interpreted to have been erupted in a rifted ocean island or island-arc environment where mantle metasomatism occurred as a result of subduction (Rajamani et al. 1985).

Many metavolcanic assemblages in the southern Abitibi greenstone belt have a large component of tholeiitic basalt, variable amounts of komatiite, rhyolite, and locally minor iron formation (e.g., 2715 Ma Kidd–Munro, 2705 Ma Kamiskotia, 2700 Ma Noranda region in Quebec, and the 2700 Ma Kinojevis North, 2705 Ma Larder Lake and Stoughton–Roquemaure). In some of these assemblages (e.g., Kidd–Munro and Kamiskotia), icelandite and high-silica rhyolite occur (Lesher et al. 1986; Barrie 1990a). The icelandite sills and flows and high-silica rhyolites represent the fractionated component of the suite consisting of tholeiitic basalt, icelandite and high-silica rhyolite. In Phanerozoic tectonic settings, this chemically distinctive suite occurs in rift-type settings (e.g., Barrie 1990a and references therein). The presence of komatiite in these assemblages is

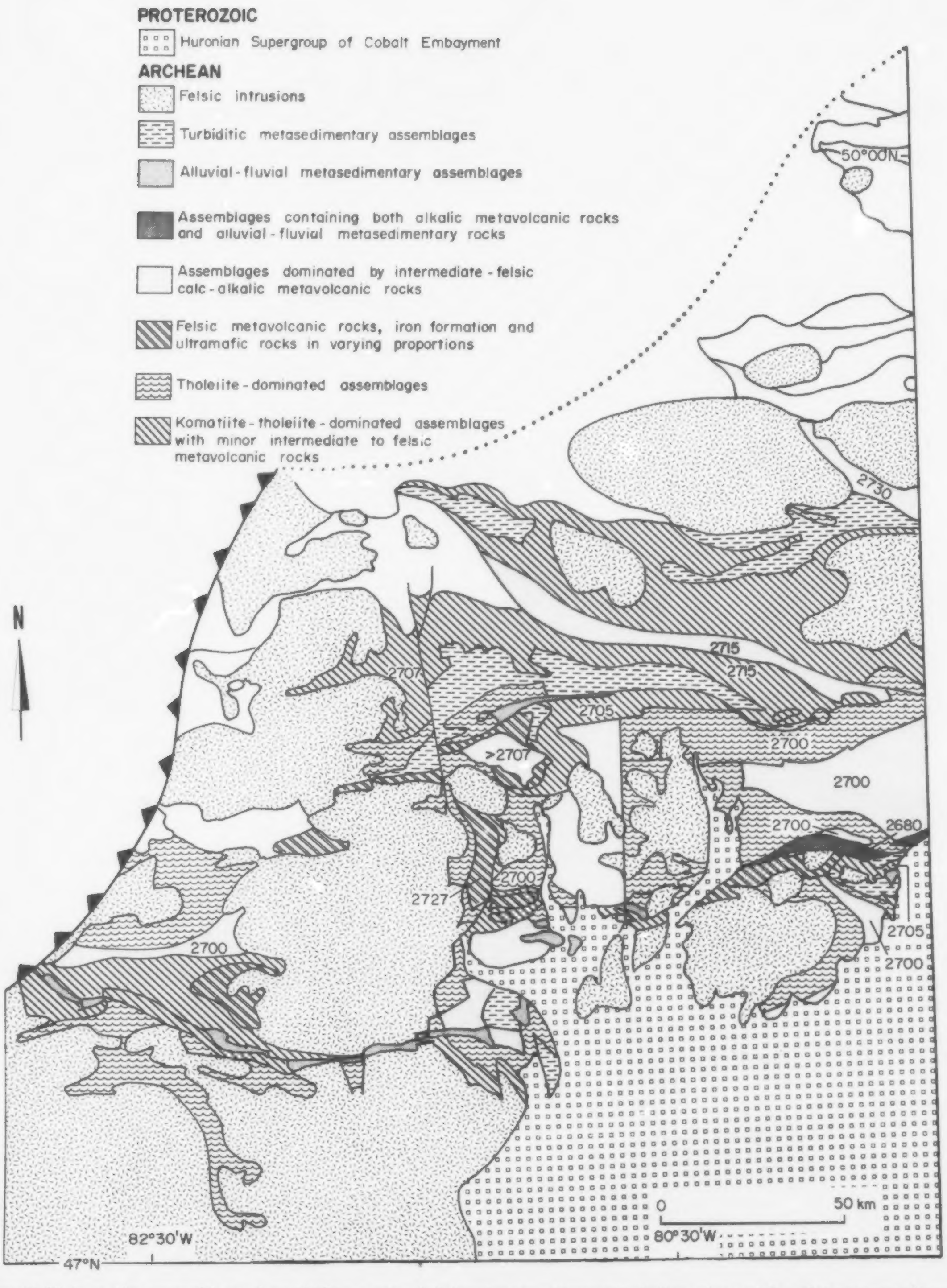
PROTEROZOIC
Huronian Supergroup of Cobalt Embayment
ARCHEAN
Felsic intrusions
Turbiditic metasedimentary assemblages
Alluvial-fluvial metasedimentary assemblages
Assemblages containing both alkalic metavolcanic rocks and alluvial-fluvial metasedimentary rocks
Assemblages dominated by intermediate-felsic calc-alkalic metavolcanic rocks
Felsic metavolcanic rocks, iron formation and ultramafic rocks in varying proportions
Tholeiite-dominated assemblages
Komatiite-tholeiite-dominated assemblages with minor intermediate to felsic metavolcanic rocks
N
50°00'N
2730
2715
2715
2707
2705
>2707
2700
2700
2700
2680
2700
2727
2705
2700
2700
0
50 km
82°30'W
80°30'W
47°N

Figure 11.35. Assemblage map of the southern Abitibi greenstone belt denoting assemblages with similar rock associations that are assumed to represent similar paleotectonic settings. (Ages shown in Ma.)

peculiar to the Archean and may reflect unique prevailing thermal conditions and chemical nature of the mantle (e.g., Ghomshei et al. 1990 and references therein). Perhaps the best documented example of a rift-like assemblage is the Kamiskotia assemblage for which Barrie (1990a) and Barrie and Davis (1990) used REE, $\varepsilon_{Nd}^{(t)}$ values, geochronology and associated base metal deposits to constrain the environment.

With the exception of the Munro Township komatiites, the geochemistry of komatiites from the Abitibi Subprovince is not sufficiently characterized to allow a comparison with komatiites from the Kolar Schist Belt or with Phanerozoic tectonic settings. However, those komatiites which are characterized by the presence of felsic metavolcanic rocks, sulphide-bearing metasedimentary rocks and nickel-sulphide mineralization may, like the Kamiskotia assemblage, have been erupted in a rifted-arc setting.

One of the central problems in resolving the tectonic setting of the tholeiite and komatiite assemblages is the character of stratigraphic basement. In the case of the Bowman assemblage, a stratigraphic substratum (Eldorado assemblage) is possible and thus, this komatiite-tholeiite assemblage may represent a rifted arc environment. For other assemblages, such as the Stoughton–Roquemaure, and Kinojevis North and Kinojevis South assemblages, stratigraphic basement has not been documented; and as for mid-ocean ridge basalts, stratigraphic basement may never have been present.

Intermediate- to Felsic-dominated Assemblages

At least 2 tectonic regimes may be interpreted for the felsic metavolcanic rock dominated assemblages. Some of the felsic metavolcanic rock dominated assemblages apparently have geochemical characteristics comparable to calc-alkalic deposits in modern-day island-arc settings (e.g., 2700 million-year-old Skead assemblage, greater than 2707 million-year-old Deloro assemblage, part of the 2700 million-year-old Blake River assemblage and possibly the 2715 Ma Duff–Coulson–Rand assemblage).

The Blake River assemblage in Ontario also has a calc-alkalic signature and is commonly attributed to a mature island-arc setting (e.g., Dimroth et al. 1983b; Hodgson and Hamilton 1989). In the Noranda area, however, Ujike and Goodwin (1987) cited REE and trace element data that led them to conclude that the most likely tectonic environment was a "...back-arc-type spreading setting..." above a primitive mantle diapir. Ujike and Goodwin noted the similarity between the Noranda massive sulphide deposits and

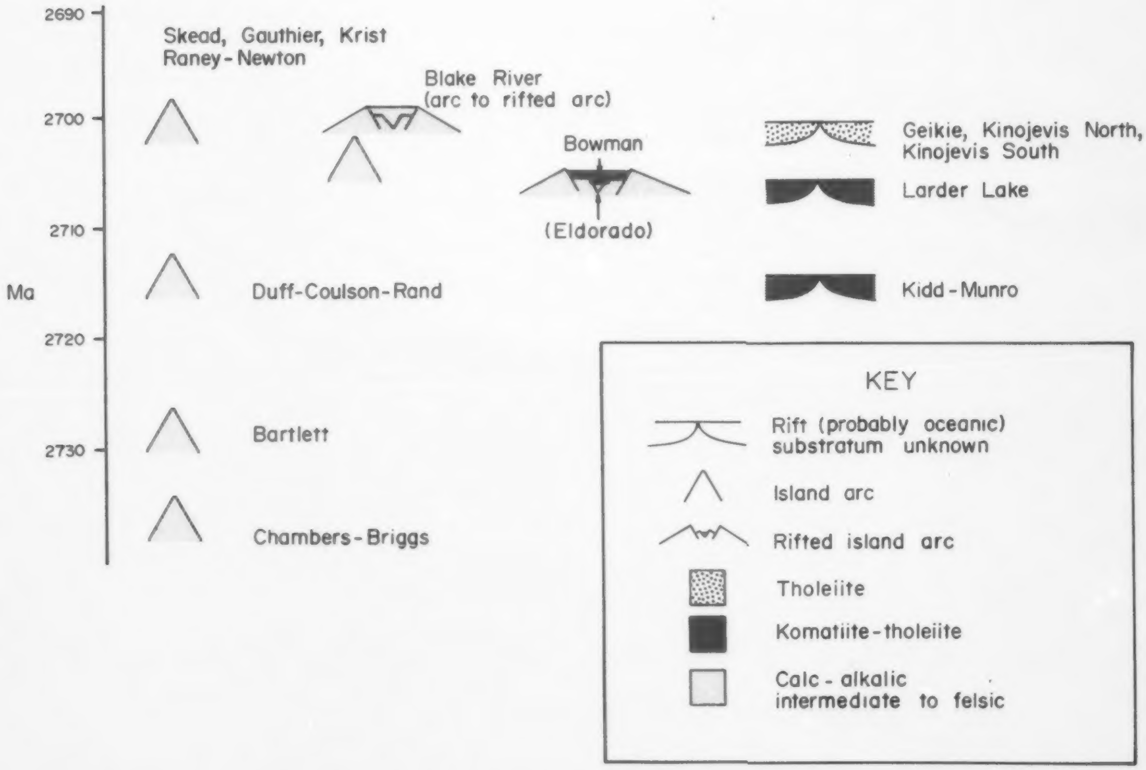
2690
2700
2710
Ma
2720
2730
Skead, Gauthier, Krist
Raney - Newton
Blake River
(arc to rifted arc)
Bowman
(Eldorado)
Geikie, Kinojevis North,
Kinojevis South
Larder Lake
Duff-Coulson-Rand
Kidd - Munro
Bartlett
Chambers - Briggs
KEY
Rift (probably oceanic)
substratum unknown
Island arc
Rifted island arc
Tholeiite
Komatiite - tholeiite
Calc - alkalic
intermediate to felsic

Figure 11.36. Time-paleotectonic chart for dated assemblages of the southern Abitibi greenstone belt. The assignment of the Bartlett assemblage to an arc environment needs to be tested; however, it does contain a significant amount of felsic volcanic rock. Similar assemblages are grouped vertically to emphasize repetition of assemblage type over time.

the Kuroko massive sulphide deposits of Japan which formed in a back-arc spreading environment. Consequently, an alternative to the commonly invoked mature island-arc setting of the Blake River assemblage (and possibly other calc-alkalic metavolcanic assemblages) is either a back-arc spreading environment or a transitional arc to back-arc spreading environment.

Assemblages containing intermediate to felsic metavolcanic rocks and ultramafic flows and/or intrusions, combined with a significant proportion of iron formation (e.g., 2727 Ma Eldorado, Bartlett and Boston assemblages) have not been geochemically characterized and consequently their environment of formation remains uncertain.

Turbidite Deposits

Deposition of at least some turbiditic metasedimentary assemblages (e.g., Porcupine and Hearst assemblages) in the southern Abitibi greenstone belt postdates ~2700 Ma; the age of other turbidite-dominated assemblages has not been established (e.g., *see* Hoyle Assemblage). In addition, some metasedimentary rocks in the Pontiac Subprovince are at least as young as 2700 Ma (Gariepy et al. 1984); although,

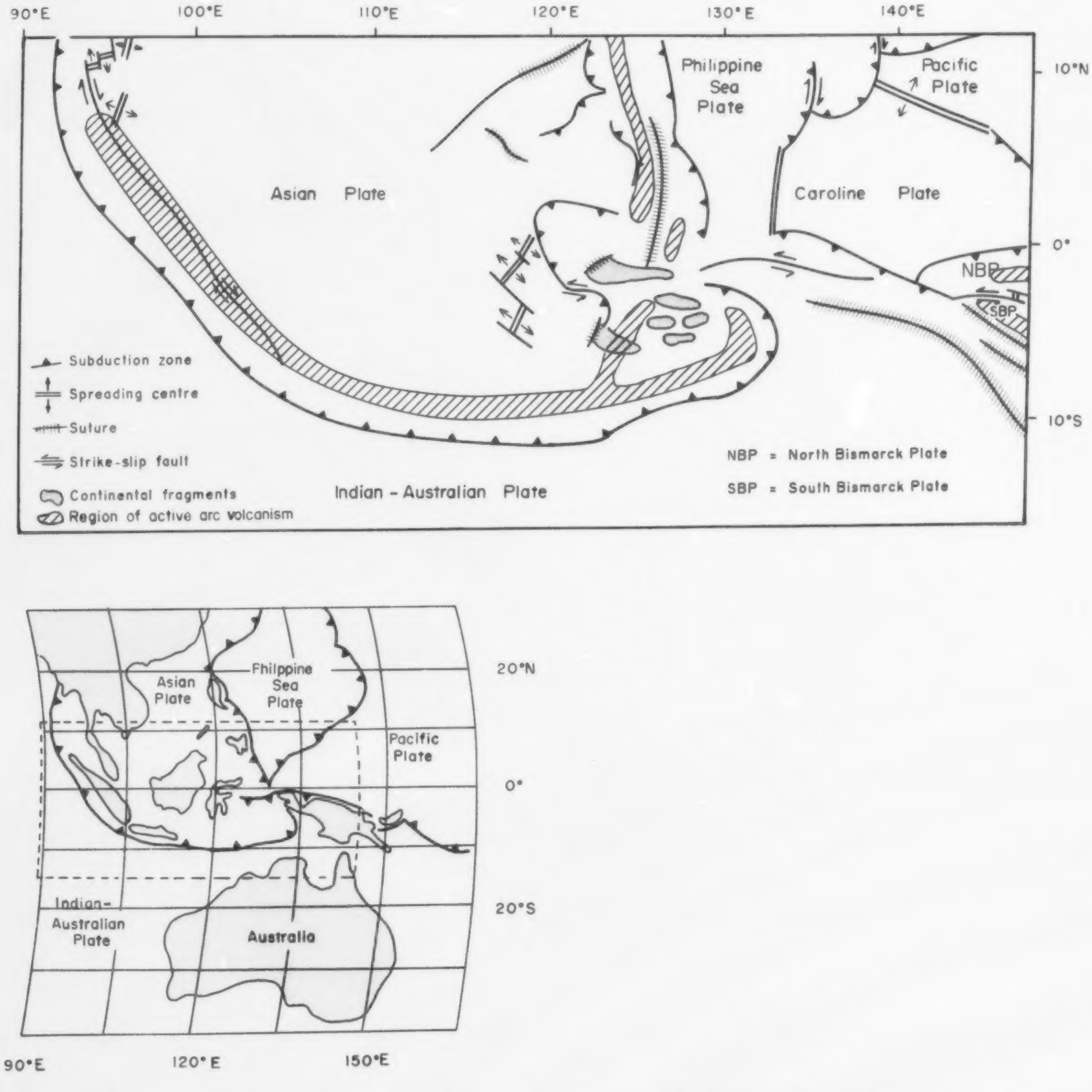
90°E
100°E
110°E
120°E
130°E
140°E
10°N
0°
10°S
Philippine Sea Plate
Pacific Plate
Asian Plate
Caroline Plate
NBP
SBP
Subduction zone
Spreading centre
Suture
Strike-slip fault
Continental fragments
Region of active arc volcanism
Indian - Australian Plate
NBP = North Bismarck Plate
SBP = South Bismarck Plate
Asian Plate
Philippine Sea Plate
Pacific Plate
20°N
0°
20°S
Indian-Australian Plate
Australia
90°E
120°E
150°E

Figure 11.37. Schematic diagram of the tectonics of the Indonesian region (*modified from* Hamilton 1988). Note the diverse tectonic environments producing different assemblages. Island-arc assemblages are being produced above subduction zones while oceanic crust is being generated at spreading centres. Some volcanism is being superimposed on both older continental fragments and on previously collided (sutured) island-arc complexes. The Indonesian region, where there is juxtaposition and superposition of diverse volcanic assemblages, may be a good analogue for the southern Abitibi greenstone belt prior to closure of ocean basins.

Table 11.2. Constraints on the contribution of mantle and crustal components to some magmatic elements of the southern Abitibi Greenstone Belt.

Magmatic Entity	Age (Ma)	Possible Crust/Mantle Contributions
Catharine assemblage (1, 2)	2720 unpub.	light REE-depleted mantle source (no evidence of evolved crust)
Kamiskotia assemblage (3)	2715	mantle depleted in radiogenic Nd (no evidence of evolved crust)
Skead assemblage (2, 4)	2700	mantle undepleted in REE (no evidence of evolved crust)
Kinojevis assemblage (1, 2)	2700 unpub.	undepleted to slightly LREE-depleted mantle (no evidence of evolved crust)
Blake River assemblage (2, 4, 5, 6, 7)	2700	mantle undepleted in REE or partial melting of tholeiitic basalt to komatiite (no evidence of evolved crust)
Timiskaming assemblage (1, 3, 8)	2685–2675 unpub.	LIL-enriched source depleted in radiogenic Nd, but less depleted than assemblages ≤ 2700 Ma and thus there could be an isotopically evolved component combined with a depleted mantle source
Granitoid "C" (3) Kamiskotia area	2695	crustal component indicated by >2926 Ma inherited zircon
Lake Abitibi Batholith (9, 10)	2690	LIL-enriched mantle; by analogy with Rainy Lake area, may be from mantle depleted in radiogenic Nd relative to rocks ≤ 2700 Ma
Otto Stock (4, 9)	2680	LIL-enriched alkalic suite that includes the Timiskaming assemblage which is depleted in radiogenic Nd relative to rocks≤2700 Ma; contribution of isotopically evolved, older component cannot be ruled out
Pressiac–Lacorne Batholith (Quebec) (11, 12, 13)	post–2700	S-type mineralogical characteristics suggest derivation by partial melting of metasediment; Pb isotope data indicates non-negligible older crustal component (~250 to 300 my older than metavolcanic rocks)
Lac Dufault Batholith (Quebec) (10, 13)	2690	Pb isotope data indicates non–negligible older crustal component (~250 to 300 my older than metavolcanic rocks)

1) Corfu, unpublished data; 2) Capdevila et al. 1982; 3) Barrie and Davis 1990; 4) Corfu et al. 1989; 5) Ujike and Goodwin 1987; 6) Fowler and Jensen 1989; 7) Smith 1980; 8) Ujike 1985; 9) Sutcliffe et al. 1990; 10) Mortensen 1987a, 1987b; 11) Dawson 1966; 12) Card et al. 1981; 13) Gariepy and Allegre 1985.

it is not clear if the turbidites of the Pontiac Subprovince are related to the turbidites in the southern Abitibi greenstone belt (e.g., Hearst and Porcupine assemblages).

At ~2700 Ma, the main period of volcanism in the southern Abitibi greenstone belt terminated, possibly due to collision with the Minnesota River Valley gneiss terrane(?), or some unknown terrane to the south, which terminated subduction. The turbidite deposits could be linked to this terminal collision since collision may have generated emergent regions subject to erosion. The turbidites then may represent "flysch"-like deposits that accumulated in submarine trenches and/or isolated distal basins at convergent plate margins. Turbidite deposits in the flysch to molasse transition documented by Crook (1989) at the southern boundary of the South Bismarck Plate (Papau New Guinea) provide a potential modern analogue for some turbidite deposits of the southern Abitibi greenstone belt.

Alkalic Metavolcanic Assemblages

The alkalic metavolcanic rocks in Timiskaming assemblages (~2685 to 2675 Ma) may be mantle derivatives; however, Timiskaming metavolcanic rocks in the Kirkland Lake area are apparently depleted in radiogenic neodymium relative to pre–2700 Ma assemblages (Barrie and Shirey 1989). Thus, the alkalic metavolcanic rocks of the Timiskaming assemblage in the Kirkland Lake area are either derived from a mantle source depleted in radiogenic neodymium relative to pre–2700 Ma assemblages, or they are derived from a mantle source comparable to the pre–2700 Ma assemblages that were contaminated by material with relatively less radiogenic neodymium material than the pre–2700 Ma assemblages (e.g., "crustal" materials). The contaminant in the later case could be in the form of subducted sediments and/or derived fluids, or in the form of crustal material. The contamination scenario warrants further examination, since there is both lead isotope (Gariepy and Allegre 1985) and inherited zircon evidence (Barrie and Davis 1990; Corfu et al., in press; *see* Table 11.2) that some post–2700 Ma intrusions contain an older "crustal" component.

The most common Phanerozoic tectonic settings of alkalic volcanic rocks are continental rift environments, island-arc environments and oceanic islands (e.g., Morrison 1980). The geochemical characteristics of the Timiskaming assemblage alkalic metavolcanic rocks are comparable to island-arc related alkalic rocks such as those of the Roman region in Italy (Cooke and Moorhouse 1969; Basu et al. 1984; Ujike 1985) and are distinct from within-plate alkalic volcanic rocks formed in either continental- or oceanic-rift environments (Ujike 1985). Some modern alkalic provinces, to which the Timiskaming metavolcanic rocks are chemically comparable, are associated with the terminal stages of subduction involving the leading edge of thinned continental crust to transitional oceanic or continental crust. Two examples include the Roman region of Italy (Ellam et al. 1989), which is associated with the leading edge of the African Plate, and the Eastern Sunda Arc in Indonesia, which is associated with the leading edge of the Australian Plate (Varne 1985). Controls on the genesis of these alkalic volcanic provinces appear related to some combination of subducted continental crust, subducted sediment, subducted subcontinental mantle and steepening and/or flipping of subduction zones (Morrison 1980; Varne 1985; Ellam et al. 1989). By analogy, then, the Timiskaming alkalic metavolcanic rocks may be related to terminal subduction involving evolved crustal material and possibly involving rapid and complex plate interactions.

PONTIAC SUBPROVINCE

The Pontiac Subprovince (*see* Figure 11.33) consists of greywacke, metasedimentary-derived anatectic granites, diorite-syenite plutons and a relatively small greenstone belt (Jolly 1978; Van de Walle 1978; Card et al. 1981; Card 1982; Dimroth et al. 1982; Lajoie and Ludden 1984). The metamorphic grade of rocks in the Pontiac Subprovince is generally in the amphibolite facies, but drops off to greenschist facies near the Larder–Cadillac shear zone (Jolly 1978). Folds within the Pontiac Subprovince verge south (Dimroth et al. 1983a). Detrital zircons indicate that the metasedimentary rocks are at least as young as 2.70 Ga and that they contain a component at least as old as 2.94 Ga (Gariepy et al. 1984). Dated granitoid rocks in the Pontiac Subprovince generally range in age from ~2695 to 2670 Ma, but there is one intrusion dated at 2632 Ma (Machado et al. 1990). Monazite and sphene from the granitoid rocks range in age from 2660 to 2640 Ma (Machado et al. 1990). The Pontiac metasedimentary rocks had a 2700 million-year-old detrital component and were intruded at 2695 Ma and later; therefore, deformation shortly followed deposition. The presence of 2695 million-year-old intrusions in the Pontiac Subprovince (Machado et al. 1990) indicates that Pontiac deposition began prior to deposition of Timiskaming-like metasedimentary rocks that are 2685 to 2675 Ma. Sources for the 2.9 billion-year-old detrital zircons could be the Minnesota River Valley gneiss terrane or old 2.9 billion-year-old rocks of the Wawa Subprovince (*see* Williams et al., this volume). Finally, no ages greater than 2.7 Ga have been found for any granitoid rocks in the Pontiac Subprovince (Machado et al. 1990); indeed the oldest rocks appear to be the metasedimentary rocks themselves.

RAMSEY–ALGOMA GRANITOID COMPLEX

The Ramsey–Algoma granitoid complex (*see* Figure 11.33) consists largely of a massive to foliated granite-granodiorite suite that intrudes a tonalite-granodiorite suite (Card 1979). Metamorphic grade of supracrustal remnants and greenstone belts is generally amphibolite facies. Only 2 ages for units from the Ramsey–Algoma gneiss terrane are available. Adjacent to the Batchawana greenstone belt, a tonalite unit has been dated at 2716±2 Ma (Corfu and Grunsky 1987). The granulite facies Levack gneisses, located just north of the Sudbury Structure, have an approximate age of 2711±7 Ma and a metamorphic age from a pegmatite of 2647±2 Ma (Krogh et al. 1984).

RELATIONSHIPS BETWEEN THE SOUTHERN ABITIBI GREENSTONE BELT, RAMSEY–ALGOMA GRANITOID COMPLEX AND THE PONTIAC SUBPROVINCE

Granitoid rocks and metavolcanic enclaves within the Ramsey–Algoma granitoid complex are similar and apparently coeval with those of the southern Abitibi greenstone belt. The greater proportion of granitoid to supracrustal rocks in the region between the Ivanhoe Lake cataclastic zone (*see* Figure 11.33) and the faults that extend northwards from, and mark the western edge of, the Cobalt Embayment (*see* Figure 11.33) may be a result of a deeper erosion level in this region.

Proterozoic cover of the Cobalt Embayment (*see* Figure 11.33) obscures the junction of the Pontiac Subprovince with the southern Abitibi greenstone belt and the Ramsey–Algoma granitoid complex. Archean outliers (e.g., Temagami area) within the Cobalt Embayment consist largely of granite-greenstone assemblages and suggest a similarity, at least at the present erosion level, to the Abitibi greenstone belt. This suggests that the southern Abitibi greenstone belt and/or Ramsey–Algoma granitoid complex extends under much of the Proterozoic cover rocks. A regional positive gravity anomaly that extends from Englehart to Sudbury to Elliot Lake (Gupta and Grant 1985) may be a manifestation of the crustal suture between the Abitibi Subprovince and the Pontiac Subprovince. The surface trace of the boundary between Pontiac units and Abitibi units, although close to, does not and need not correspond to the trace of the gravity anomaly.

SUMMARY OF TECTONIC MODELS

Megacauldron Model

Jensen and Langford (1985 and references therein) proposed a "megacauldron" model for the Abitibi greenstone belt in which metavolcanic accumulations developed above a mantle diapir. Upon adiabatic decompression, the diapir undergoes partial melting resulting in eruption of komatiitic-tholeiitic metavolcanic rocks on a crust of unspecified but presumably mafic-ultramafic composition. The resultant loading of the crust is envisaged to initiate crustal-scale subsidence of such an extent that the base of the mafic-ultramafic pile is converted to eclogite, which partly melts to form calc-alkalic metavolcanic rocks. Continuing subsidence results in higher-grade metamorphism and partial melting of the calc-alkalic metavolcanic rocks to form trondhjemite plutons. The alkalic metavolcanic rocks are presumed to be derived from the metasedimentary rocks that developed at the margins of the megacauldrons. The apparent cyclic volcanism within the Abitibi greenstone belt is explained by successive megacauldrons and overlapping of deposits from adjacent megacauldrons.

Although mantle diapirism and the megacauldron model developed by Jensen (1985a) may account for the basic petrogenetic features of the pre–2700 Ma metavolcanic assemblages, there are 3 major concerns with this model:

1. It does not explain the recent discovery that some post–2700 Ma intrusions appear to have a crustal component within them that is up to 300 my older than the oldest known metavolcanic rocks (e.g., Gariepy and Allegre 1985).
2. It cannot explain the south-verging folds in the Pontiac Subprovince and the higher metamorphic grade of the Pontiac metasedimentary rocks that are in part as young as 2700 Ma.

Ensialic Rift Model

Hodgson (1986) suggested that the southern Abitibi greenstone belt formed on pre-existing sialic crust during extensive rifting and mantle diapirism that progressed from ensialic to ensimatic rifting. The first-formed volcanic products are presumed to be komatiitic to tholeiitic in composition erupted from mantle diapirs. Calc-alkalic metavolcanic rocks were presumed to be formed by partial melting of the sialic crust. The extensive rifting and successive overprinting of normal faults was used to explain how a large apparent thickness of steeply dipping metavolcanic rocks could occur without penetrative deformation. The regional shear zones were assumed to be early formed listric faults that were reactivated during tectonism. Regional deformation was assumed to result from both squeezing of metavolcanic strata between rising diapirs and crustal shortening. Hodgson (1986) suggested that the Timiskaming assemblage formed during this compressional phase. The 2 main objections to this model are outlined below:

1. There is no evidence for an older sialic stratigraphic basement to the southern Abitibi greenstone belt metavolcanic rocks.
2. The basic stratigraphy and geochemistry of the southern Abitibi greenstone belt does not support an ensialic rift environment (Jackson and Sutcliffe 1990; *see* Table 11.2). No continental rift clastic sediments or continental rift volcanic rocks are reported among the oldest supracrustal rocks of the southern Abitibi greenstone belt.

Island-arc Models

Three models outlined below are variations on a theme of island-arc tectonics as first formulated by Dimroth et al. (1983b). In detail, however, the models differ dramatically. For example, the Pontiac Subprovince is envisaged as a fore-arc basin (Dimroth et al. 1983b), a foreland to a continental mass to the south (Hodgson and Hamilton 1989) and a "deep-water rift basin" in the back-arc (Ludden et al. 1986). The basic features of the island-arc models are summarized below.

Dimroth et al. (1982, 1983a, 1983b) on the basis of detailed mapping and regional synthesis proposed an island-arc model for the southern Abitibi greenstone belt, wherein the arc presumably formed over a north-dipping subduction zone. In this model, the ultramafic to mafic tholeiitic assemblages were interpreted as an oceanic platform (crust) to the calc-alkalic central metavolcanic complexes. Coeval with the calc-alkalic volcanism, Dimroth et al. (1983b) envisaged that accretion of large volumes of "granite" to the base of the arc assemblage facilitated uplift and erosion of the arc complex. Maturing of the island arc resulted in development of a metasedimentary basin (accretionary wedge?) to the south. The metasedimentary Pontiac Subprovince terrain was in part overthrust by the more northerly island-arc assemblage during protracted north-south compression. Collision with an unspecified terrain to the south terminated subduction.

The tectonic model proposed by Dimroth et al. (1982, 1983a, 1983b) is attractive in that it is complete from genesis of the metavolcanic rocks to their tectonism in one simple evolutionary plate-tectonic regime. Furthermore, it provides a mechanism for contaminating some of the late magmatic elements of the southern Abitibi greenstone belt with evolved crustal material (e.g., Gariepy and Allegre 1985; Jackson and Sutcliffe 1990, and references therein). It also provides a mechanism, tectonic loading, to metamorphose the Pontiac Subprovince in which deposition of sedimentary rocks began *circa* 2700 to 2695 Ma.

Ludden et al. (1986) provided an alternate view of the Abitibi greenstone belt which involved development of a continental-arc complex (the northern Abitibi greenstone belt) prior to 2720 Ma, followed by rifting of this older arc between 2710 and 2700 Ma to form the southern Abitibi greenstone belt. The arc and rifted-arc complexes were postulated to develop above a south-dipping subduction zone. Wrench faulting (Hubert et al. 1984) and other unspecified orogenic processes are postulated to account for the present structural complexity of the belt.

Hodgson and Hamilton (1989) expanded and modified the island-arc model of Dimroth et al. (1983b). In their model, north-dipping subduction formed island-arc assemblages on pre-existing oceanic crustal assemblages. Subduction led to terminal collision with the foreland basin of the leading edge of a southerly continental mass, now the Pontiac Subprovince. Collision involved the southward overthrusting of the southern Abitibi greenstone belt over the Pontiac Subprovince and synchronous development of molasse deposits, the Timiskaming group.

The preceding plate-tectonic models assume some variation of a single evolving arc-trench system involving some collisional process. The repetition of similar assemblages at different times (e.g., the rift-like (arc and/or oceanic) assemblages at 2715, 2705 and 2700 Ma; *see* Figure 11.36) suggests that similar tectonic processes were operative at different times. The juxtaposition of coeval assemblages, that are dramatically different in terms of rock association, suggests amalgamation of assemblages that represent different but contemporaneous tectonic environments. These 2 inferences (repetition of the same tectonic process at different times, and different tectonic processes operative at the same time) suggest a tectonic history much more complicated than a simple evolving single arc-trench system. A potential analogue for the southern Abitibi greenstone belt is the Indonesian region (e.g., Hamilton 1978, 1979; *see* Figure 11.37). In the absence of a better understanding of the environment of formation of the Abitibi Subprovince metavolcanic assemblages, the analogy to Indonesia cannot be carried much further at this time. In the Indonesian region, there are contemporaneous but spatially separated arcs and spreading centres (coeval but differing metavolcanic assemblages of the southern Abitibi greenstone belt?). There are also older island-arc and ophiolite complexes that have been tectonically assembled and affected by younger superimposed tectonic products (similar assemblages of different ages in the southern Abitibi greenstone belt?; Hamilton 1978, 1979).

The Minnesota River Valley gneiss terrane (*see* Figure 11.33), which consists in part of 2.8 to 3.5 Ga gneisses (Sims et al. 1980), may have been the mass involved in terminal collision with the southern Abitibi greenstone belt (e.g., Hoffman 1989 and references therein). This old gneiss terrane is located roughly 1000 km to the west of the Abitibi Subprovince. In Late Archean time, it may have extended farther to the east and was then subsequently obscured and/or removed in Proterozoic–Phanerozoic time by sedimentary cover and Proterozoic orogens (e.g., Penokean and Grenville). There are several relationships consistent with the hypothesis that the Minnesota River Valley gneiss terrane was involved in a collisional event in the late Archean.

1. The Great Lakes Tectonic Zone (*see* Figure 11.33) is an ~30° north-dipping, as interpreted using seismic reflection (Gibbs et al. 1984), linear zone of high strain, tight folding and faulting, that separates a 2.75 to 2.6 billion-year-old granite-greenstone assemblage to the north from the much older (2.8 to 3.5 Ga) gneiss terrane to the south (Sims et al. 1980; Morey et al. 1982).
2. The gneiss terrane experienced at least 2 periods of high-grade metamorphism at 3.05 Ga and ~2.6 Ga (e.g., Goldich and Wooden 1980; Goldich et al. 1980). The ~2.6 billion-year-old metamorphism is roughly coeval with the Late Archean suturing of the granite-greenstone terrane and the gneiss terrane along the Great Lakes Tectonic Zone. In addition, this ~2.6 Ga age is also similar to "young" ages of deep crustal metamorphism in the Superior Province of Canada: Ryan Migmatite at ~2.62 Ga (Krogh and Turek 1982); gneisses of the Kapuskasing Structural Zone at ~2.65 to 2.63 Ga (Corfu 1987; Percival and Krogh 1983); Levack gneisses at ~2.65 Ga (Krogh et al. 1984); and ~2.66 to 2.64 Ga in the Pontiac Subprovince (Machado et al. 1990).

SUMMARY

Early formed volcanic assemblages accumulated between 2.75 and 2.70 Ga in a variety of coeval and repeating tectonic environments; perhaps in a tectonic setting comparable to the Indonesian region which displays numerous diverse microplate interactions. Metavolcanic rocks between 2750 and 2700 Ma formed in subaqueous settings; clastic sediment formation, concentration and subsequent deposition were not major processes prior to 2700 Ma, indicating that any pre–2700 Ma trenches were possibly starved of sediment. It is assumed that the region of microplate interaction was trapped between 2 larger convergent masses, the Minnesota River Valley gneiss terrane to the south and the Uchi Subprovince and previously accreted material to the north. Approximately 2700 Ma, terminal collision began, accreting and deforming the microplate region. During the collision, it is postulated that large areas in the collisional region became emergent, thereby facilitating erosion, collection, transport and deposition of sediment. As a result, large turbidite deposits formed, including either a contribution to, or the entire formation of, the Pontiac Subprovince. Thus, the Pontiac Subprovince may represent a rapidly formed, 2.70 to 2.69 billion-year-old accretionary wedge of turbiditic deposits. Other turbidite basins may have formed locally at this time in the still submerged distal parts of the orogen. During the postulated collision, the southern Abitibi greenstone belt was thrust over the Pontiac Subprovince. The collision terminated subduction and thus volcanism, with the exception of the alkalic metavolcanic rocks at 2.69

to 2.68 Ga. These alkalic metavolcanic rocks represent the terminal stages of subduction and may be related to complex interactions of lithospheric plates during the terminal stages of subduction, subduction of sediment and underthrusting of evolved continental material. The alluvial-fluvial facies of the Timiskaming and Timiskaming-like assemblages formed during the late stages of tectonism synchronous with the development of the regional shear zones and may represent a mature arc or "continental" environment. Near the Larder–Cadillac shear zone, these assemblages may record the final suturing of allochthonous terranes of the southern Abitibi greenstone belt and the inferred, more southerly continental mass. There is, however, much controversy surrounding the tectonic significance of the shear zones and the Timiskaming metavolcanic and metasedimentary rocks, e.g., molasse (Hodgson and Hamilton 1989) versus strike-slip basin deposits (e.g., Colvine et al. 1988; Thurston and Chivers 1990).

A better understanding of the supracrustal assemblages that make up the southern Abitibi greenstone belt, in particular their ages, internal structure, boundary relationships and geochemical characteristics, and better characterization of regional relationships of subprovince-scale entities in the southern Superior Province will lead to development of better constrained evolutionary models for the southern Abitibi greenstone belt.

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Chapter 12

Wawa Subprovince

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Abstract

In the Wawa Subprovince, at least 3 stages of supracrustal development are preserved, dating from approximately 2.89, 2.75 and 2.70 Ga. Tectonic assemblages of komatiitic to tholeiitic basalt are overlain by iron- and magnesium-rich tholeiite, and by sporadic centres of submarine to subaerial calc-alkalic basalt and intermediate to felsic volcanic rocks. Volcanic sequences are commonly overlain by, or are sources of and transitional into, aprons of sediment. Some assemblages developed one upon another, others formed separately and became tectonically juxtaposed to form aggregates, called greenstone belts. Unconformities between assemblages are uncommon; most assemblages were juxtaposed along shallow-dipping shear zones and some were subjected to recumbent folding.

Between these greenstone belts, that form but 20 to 30% of the subprovince, are granitoid rocks that consist of massive, foliated and gneissic tonalite-granodiorite which are cut by massive to foliated granodiorite and granite. Most granitoids were emplaced during or after production of the supracrustal rocks with which the granitoids are spatially associated. Granitoid emplacement was in part synchronous with upright folding and controlled the shapes and sizes of the characteristically steeply dipping supracrustal belts.

The Wawa Subprovince was amalgamated with the Quetico accretionary prism to the north by processes caused by plate convergence and transpressive interaction during the late Neoarchean era. Pull-apart basins and spatially associated alkaline volcanism developed as a result of continued plate convergence after subprovince amalgamation. The subprovince is the site of iron and gold deposits.

INTRODUCTION

The Wawa Subprovince is an aggregation of Archean greenstone belts and granitoid plutons (Figure 12.1). The northern boundary (Figure 12.2) lies in tectonic contact against structurally overlying metasedimentary rocks of the Quetico Subprovince (*see* Williams, this volume). The southern boundary of the Wawa Subprovince is marked by the Montreal River fault in the east, and is hidden beneath Lake Superior and by unconformable Proterozoic supracrustal rocks of the Animikie Basin in the west.

The Great Lakes Tectonic Zone separates the Wawa Subprovince from the Marquette greenstone belt and the Minnesota River Valley gneiss terrane to the south (Sims et al. 1980). The subprovince extends from the Kapuskasing Structural Zone in the east where the Wawa Subprovince is interpreted by Percival and Card (1985) to be transitional into the Chapleau block and Val Rita block, forming the southern and central parts of the Kapuskasing Structural Zone. The western end of the Wawa Subprovince abuts against the Proterozoic Trans–Hudson Orogen, which trends southward through Manitoba; both are hidden beneath Phanerozoic rocks in North Dakota.

The Wawa Subprovince is composed of 2 linear concentrations of greenstone belts; one along its northern border with the Quetico Subprovince, and another in the Mishibishu–Michipicoten–Gamitagama area. These broad zones, enriched in supracrustal rocks, are separated by belt-like domains of plutonic rocks (*see* Figure 12.2). Regional stratigraphic correlations are not fully developed at present and geochronologic investigations are concentrated only within the Michipicoten, Wawa (Turek et al. 1984; Turek, Keller and Van Schmus 1990), Hemlo (Corfu and Muir 1989a) and Shebandowan greenstone belts (Corfu and Stott 1986). Accordingly, this chapter is subdivided into sections that cover, from southeast to northwest: the greenstone belts of Gamitagama, Mishibishu and Michipicoten; the Dayohessarah–Kabinakagami, Manitouwadge–Hornepayne and Schreiber–Hemlo greenstone belts; the 2 major plutonic areas of the Pukaskwa and Black–Pic batholiths; and the Shebandowan greenstone belt in the west (*see* Figure 12.2). For each greenstone belt, descriptions of supracrustal and plutonic rocks, their structures and mineral deposits are followed with an interpretation of the sequence of tectonic events. Concluding the chapter, there is a brief review and discussion of this information, a proposal for a tectonic sequence of events and a comparison of the development of the Wawa Subprovince with that of the neighbouring Quetico and Abitibi subprovinces.

There is a considerable disparity in the amount of geological information available in different parts of this subprovince and this is reflected in the variable detail in the descriptions. All Archean supracrustal rocks are metamorphosed and the prefix *meta-* (in metavolcanic and metasedimentary rocks) has generally been deleted to ease readability in this chapter.

The Wawa Subprovince is a granite-greenstone terrane in which disparate units, and well-defined greenstone belts of metamorphosed komatiite, basalt, dacite and rhyolite and associated metasedimentary rocks, are dispersed in a sea of granitoid rocks (*see* Figure 12.2). The metasedimentary rocks include turbiditic wacke, minor conglomerate and

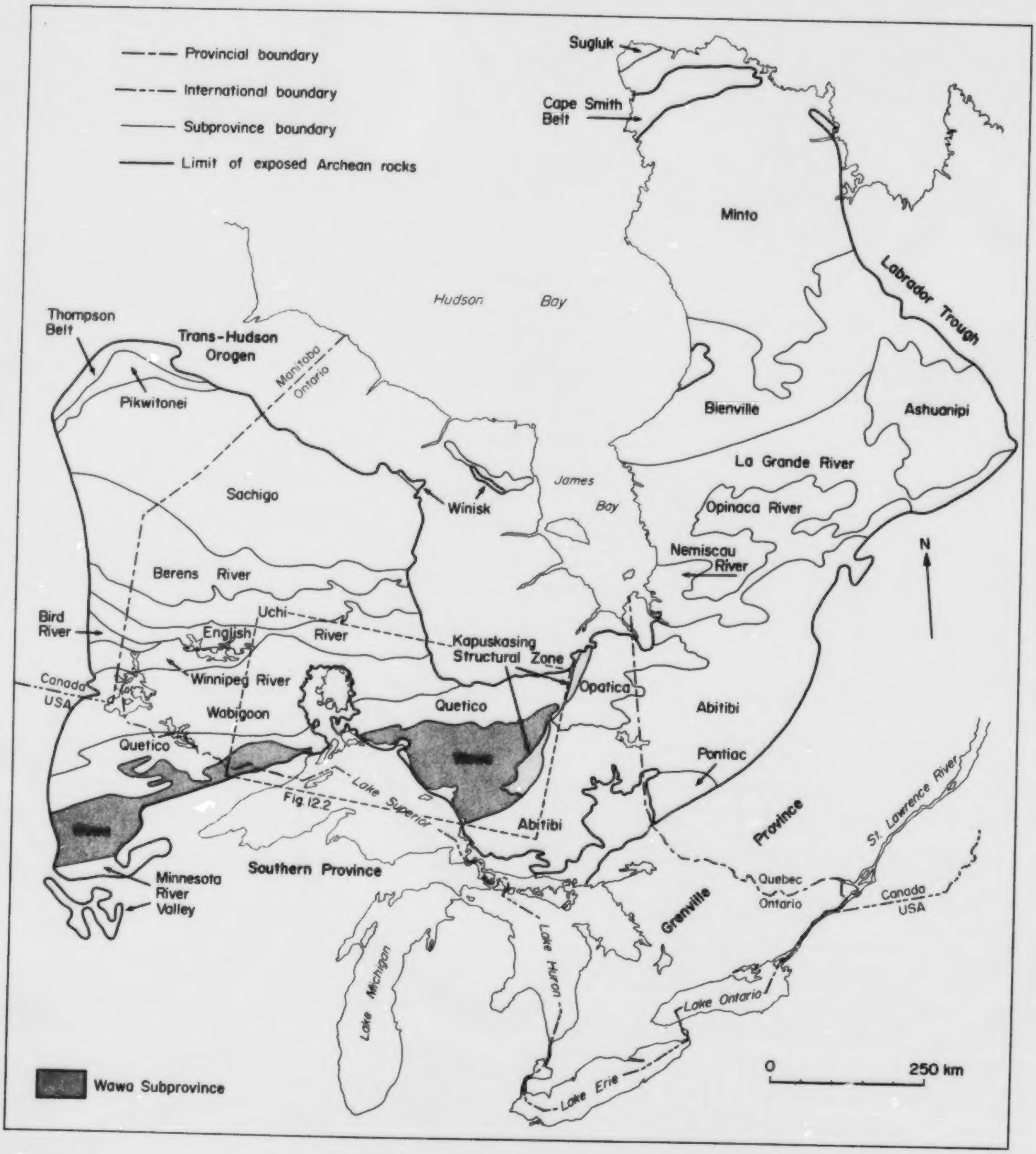

Provincial boundary
International boundary
Subprovince boundary
Limit of exposed Archean rocks
Sugluk
Cape Smith Belt
Minto
Labrador Trough
Hudson Bay
Thompson Belt
Trans-Hudson Orogen
Pikwitonei
Manitoba
Ontario
Bienville
Ashuanipi
La Grande River
James Bay
Opinaca River
Sachigo
Winisk
Nemiscau River
N
Berens River
Bird River
Uchi-
English
River
Kapuskasing Structural Zone
Canada
USA
Winnipeg River
Wabigoon
Quetico
Opatica
Abitibi
Quetico
Wawa
Pontiac
Fig.12.2
Lake Superior
Province
St. Lawrence River
Wawa
Abitibi
Southern Province
Minnesota River Valley
Quebec
Ontario
Canada
USA
Grenville
Lake Michigan
Lake Huron
Lake Ontario
0
250 km
Lake Erie
Wawa Subprovince

Figure 12.1. Location map showing the Wawa Subprovince, neighbouring Quetico Subprovince, and the Kapuskasing Structural Zone (*modified from* Card and Ciesielski 1986).

iron formation. Stratigraphic and structural relationships between these units of volcanic and sedimentary rocks are usually unclear and commonly masked by later shearing.

Rather than describe the geology of these belts in terms of simple lithostratigraphic sequences, small, homogeneous units, called assemblages, are delineated in which stratigraphic, structural, metamorphic or geochronologic consistency has been recognized or confidently interpreted. For convenience only, where assemblages are discrete but probably correlative with others, individual segments of an assemblage have been defined.

Aeromagnetic, gravity and airborne radiometric data have been accumulated and consulted. Gravity (*see* Maps 2593 and 2594, map case) and magnetic maps (*see* Maps 2585 and 2586, map case) are included with this volume. Magnetic anomalies have been partly successful in delineating structural trends of lithologic units composed of mafic volcanic rocks and iron formation. For example, along the northern edge of the Wawa Subprovince, in the Manitouwadge region, gneisses with abundant inclusions of these rock types are matched by strongly developed curvilinear magnetic trends. Unfortunately, much of the eastern part of the Wawa Subprovince has been invaded by diabase dike swarms (*see* Osmani, this volume), which have all but obliterated the Archean magnetic signature. Magmatic bodies such as the Fourbay pluton (Milne 1968) are clearly distinguished by a magnetic signature. In the western part of the Wawa Subprovince, a general distinction of granitoid-dominated regions from those rich in greenstone is possible. Bouguer gravity maps reveal the distribution of granitoid-dominated crust, characterized by low Bouguer gravity values; however, the distinction of major structures and one rock type from another has yet to be generally made using the gravity information.

Within the Wawa Subprovince, supracrustal rocks at Michipicoten have been subjected to repeated geologic investigation since Sir William Logan described boulder conglomerates containing granitic clasts near the Doré River on the shore of Lake Superior (Logan 1847). The Michipicoten and other greenstone belts such as the Mishibishu and Gamitagama greenstone belts are well exposed in the rugged topography near Lake Superior. Other greenstone belts, such as the Manitouwadge and Dayohessarah greenstone belts, are poorly exposed where topography is low and undulating. This is typical for large areas of the Superior Province. Primary textures are well preserved in the supracrustal rocks of the larger greenstone belts, such as the Michipicoten greenstone belt (sometimes referred to as the Wawa greenstone belt), in spite of greenschist facies metamorphism and considerable deformation. The smaller greenstone belts, such as the Kabinakagami and Manitouwadge greenstone belts, rarely contain primary structures. Volcanic rocks predominate in the greenstone belts; the Michipicoten greenstone belt exhibits 3 discrete cycles (Goodwin 1962); the Schreiber–Hemlo greenstone belt, only 2 cycles (Corfu and Muir 1989a). Sedimentary rocks appear to be largely derived by erosion of the felsic part of the youngest volcanic cycle in each belt, but evidence will be presented for at least 3 stages of sedimentation. Research into the rocks of these greenstone belts has been stimulated through the years by interest in both the gold and iron deposits contained within them.

The Wawa Subprovince (*see* Figure 12.2) can conveniently be subdivided into a number of greenstone belts and these greenstone belts may be further subdivided into assemblages. The assemblages are listed in Tables 12.1, 12.2 and 12.3. The greenstone belts have many lithologic and structural features in common, as well as several that distinguish them. All of these belts are Archean and some have been modified by Proterozoic tectonism and magmatism, which included the development of northwest- and northeast-striking faults and extensive diabase dikes. This chapter will deal mainly with the Archean geology of these belts. All U-Pb ages cited have been determined with high precision on zircon or sphene.

The following sections describe each greenstone belt within the Wawa Subprovince (*see* Table 12.1) and propose and delineate assemblages and their components. This is followed by a description of the structure and plutonic igneous development of each greenstone belt or of its assemblages (*see* Table 12.2). In conclusion, some general comments are made upon the structural geology, tectonic and magmatic development of the entire Wawa Subprovince.

Granitoid plutonic rocks within the Wawa Subprovince engulf the greenstone belts and occur as intrusions within them (*see* Figure 12.2). There has been little study on much of the granitoid rock between greenstone belts, but it is known to be composed of tonalite to granodiorite (e.g., Pukaskwa batholith, Black–Pic batholith, Northern Light–Perching Gull Lakes batholitic complex), and is commonly foliated to gneissic and intruded by masses of more potassic, granitic rock .

GREENSTONE BELTS AND THEIR ASSEMBLAGES

Gamitagama Greenstone Belt

The Gamitagama greenstone belt (Ayres 1969a, 1969b; Choudhry 1981) is located 45 km south of Wawa and is geographically separated from the Mishiwawa part of the Michipicoten greenstone belt by only 3 km (*see* Figures 12.2 and 12.3). Located on the east shore of Lake Superior, this small belt is approximately 26 km in length and 24 km in width and can be subdivided into a sedimentary rock-dominated assemblage and a volcanic rock-dominated assemblage (*see* Figures 12.2 and 12.3; *see also* Tables 12.1 and 12.2). Structural and geochronologic data are insufficient to support further subdivision of the volcanic assemblage, although Ayres (1969a, 1969b) did attempt to unravel the volcanic stratigraphy.

GAMITAGAMA ASSEMBLAGE

The Gamitagama assemblage occupies the northern two-thirds of the Gamitagama greenstone belt and is dominated by mafic, felsic and minor intermediate volcanic

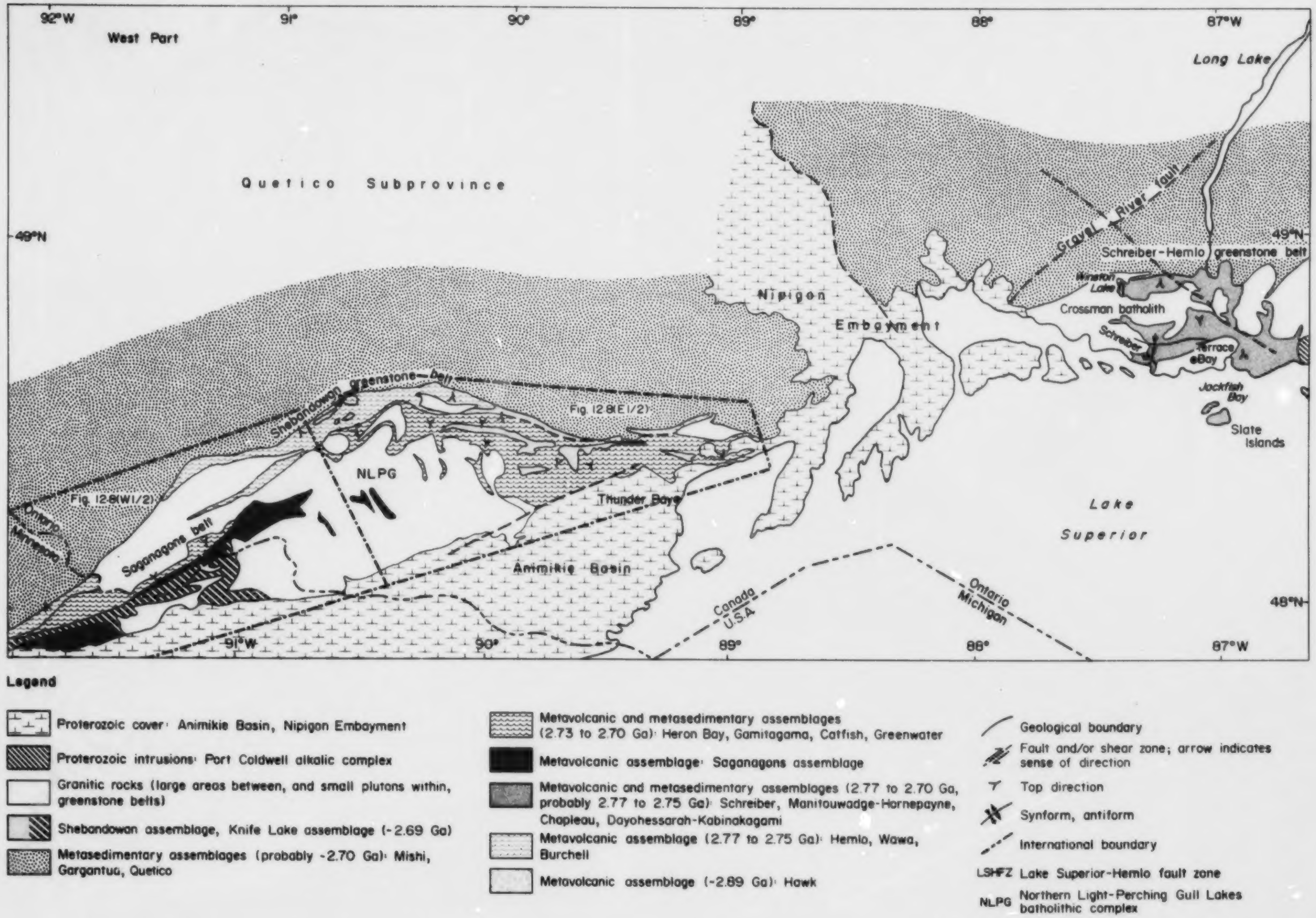

92°W
91°
90°
89°
88°
87°W
West Part
Long Lake
Quetico Subprovince
49°N
Gravel River fault
Schreiber-Hemlo greenstone belt
Winston Lake
Crossman batholith
Schreiber
Terrace Bay
Jackfish Bay
Slate Islands
Nipigon
Embayment
Shebandowan greenstone belt
Fig. 12.8(E1/2)
Fig. 12.8(W1/2)
NLPG
Ontario
Minnesota
Saganagons belt
Thunder Bay
Animikie Basin
Lake Superior
Canada
U.S.A.
Ontario
Michigan
48°N
91°W
90°
89°
88°
87°W
Legend
Proterozoic cover: Animikie Basin, Nipigon Embayment
Proterozoic intrusions: Port Coldwell alkalic complex
Granitic rocks (large areas between, and small plutons within, greenstone belts)
Shebandowan assemblage, Knife Lake assemblage (~2.69 Ga)
Metasedimentary assemblages (probably ~2.70 Ga): Mishi, Gargantua, Quetico
Metavolcanic and metasedimentary assemblages (2.73 to 2.70 Ga): Heron Bay, Gamitagama, Catfish, Greenwater
Metavolcanic assemblage: Saganagons assemblage
Metavolcanic and metasedimentary assemblages (2.77 to 2.70 Ga, probably 2.77 to 2.75 Ga): Schreiber, Manitouwadge-Hornepayne, Chapleau, Dayohessarah-Kabinakagami
Metavolcanic assemblage (2.77 to 2.75 Ga): Hemlo, Wawa, Burchell
Metavolcanic assemblage (~2.89 Ga): Hawk
Geological boundary
Fault and/or shear zone; arrow indicates sense of direction
Top direction
Synform, antiform
International boundary
LSHFZ Lake Superior-Hemlo fault zone
NLPG Northern Light-Perching Gull Lakes batholithic complex

Figure 12.2. Tectonic assemblage map of the Wawa Subprovince, showing the greenstone belts and their constituent assemblages, selected stratigraphic-facing data, major granitoid masses and faults.

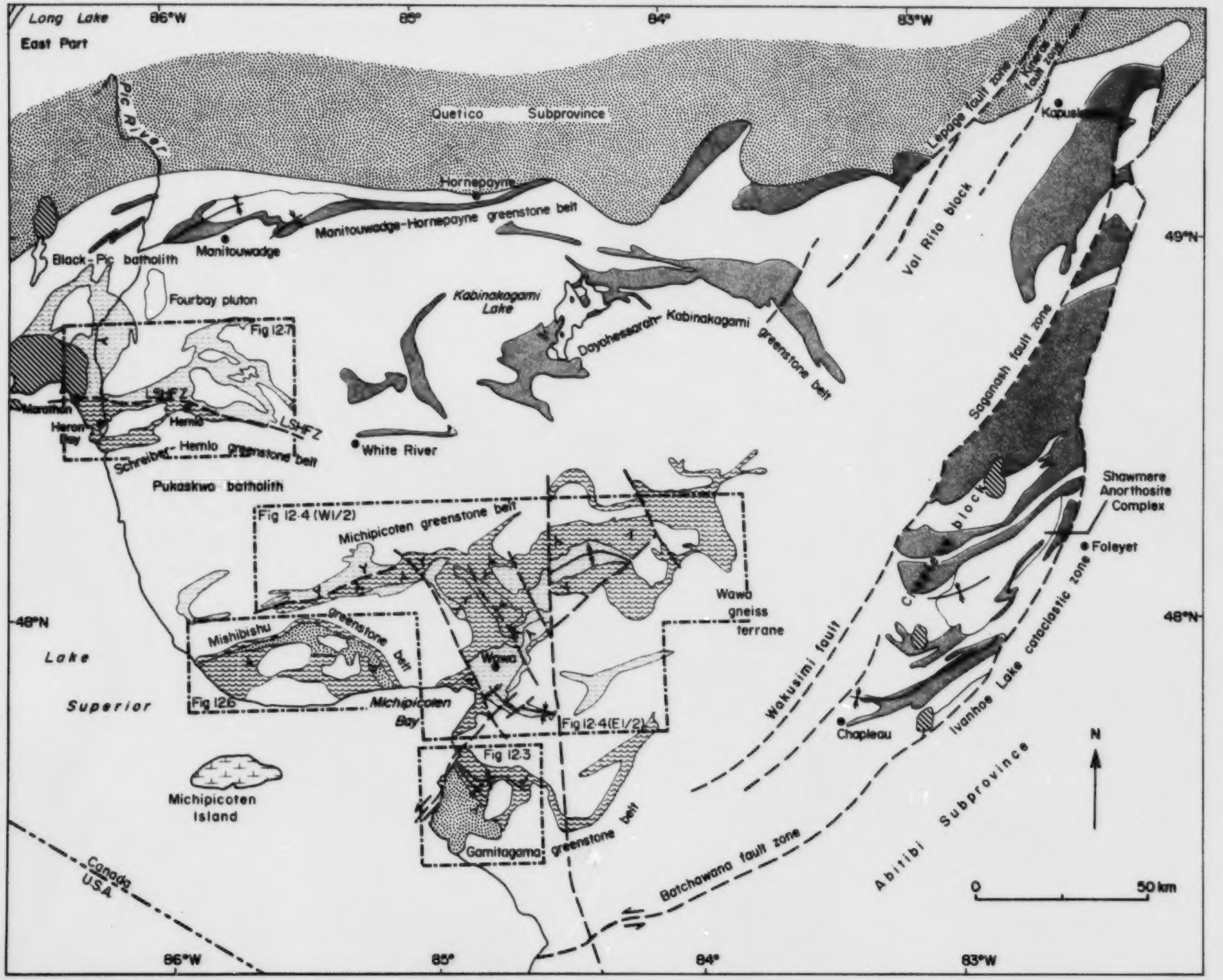

Long Lake
East Part
86°W
85°
84°
83°W
Quetico Subprovince
Pic River
Hornepayne
Manitouwadge-Hornepayne greenstone belt
Manitouwadge
Black-Pic batholith
Fourbay pluton
Kabinakagami Lake
Dayohessarah-Kabinakagami greenstone belt
Lepage fault zone
Kineras fault zone
Kapus
Val Rita block
49°N
Saganash fault zone
Fig 12·7
LSHFZ
Marathon
Heron Bay
Hemlo
Schreiber-Hemlo greenstone belt
White River
Pukaskwa batholith
Fig 12·4 (W1/2)
Michipicoten greenstone belt
block
Shawmere Anorthosite Complex
Foleyet
Wawa gneiss terrane
48°N
Mishibishu greenstone belt
Wawa
Lake Superior
Fig 12·6
Michipicoten Bay
Fig 12·4 (E1/2)
Wakusimi fault
Chapleau
Ivanhoe Lake cataclastic zone
Fig 12·3
Michipicoten Island
Gamitagama greenstone belt
Abitibi Subprovince
N
Canada
U.S.A.
Batchawana fault zone
0
50 km

Figure 12.2. Tectonic assemblage map of the Wawa Subprovince, showing the greenstone belts and their constituent assemblages, selected stratigraphic-facing data, major granitoid masses and faults.

Table 12.1. Greenstone belts in the Wawa Subprovince and tectonic assemblages therein.

Greenstone Belt	Lithotectonic Assemblages	Dominant Rock Types
Gamitagama	Gamitagama	metavolcanic rocks
	Gargantua	metasedimentary rocks
Michipicoten	Hawk	metavolcanic rocks
	Wawa	metavolcanic rocks
	Catfish	metavolcanic rocks
	Catfish segment	
	McCormick segment	
Mishibishu	Catfish	metavolcanic rocks
	Miron segment	
	Pukaskwa–Point Isacor segment	
	Mishi	metasedimentary rocks
Schreiber–Hemlo	Heron Bay	metavolcanic rocks
	Hemlo–Black River	metavolcanic rocks
	Schreiber	metavolcanic rocks
Dayohessarah–Kabinakagami	*Dayohessarah–Kabinakagami	metavolcanic rocks, minor sedimentary rocks
Chapleau block	*Chapleau	metasedimentary rocks metavolcanic rocks,
Manitouwadge–Hornepayne	*Manitouwadge–Hornepayne	metavolcanic rocks, minor sedimentary rocks
Shebandowan and Saganagons	Burchell	metavolcanic rocks
	Greenwater	metavolcanic rocks
	Saganagons	metavolcanic rocks
	Shebandowan	metasedimentary rocks
	Knife Lake	metasedimentary rocks

**unsubdivided high-grade greenstone belt*

rocks (*see* Figure 12.3). The assemblage is bounded to the north by the Rabbit Blanket Lake intrusion (2668±2 Ma; Krogh and Turek 1982) which is part of a large composite batholith of foliated to gneissic, hornblende-biotite tonalite, diorite, granodiorite and trondhjemite that separates the Gamitagama greenstone belt from the Michipicoten greenstone belt to the north (*see* Figure 12.3). The assemblage is bounded to the south by sedimentary rocks of the Gargantua assemblage and locally by granitoid intrusions (*see* Figure 12.3).

The Gamitagama assemblage is dominated by tholeiitic mafic to andesitic volcanic rocks with lesser amounts of calc-alkalic, rhyolitic to dacitic, felsic volcanic rocks. The mafic rocks consist of flows that are locally pillowed and locally feldspar porphyritic (Ayres 1969b). Gabbro and diorite intrusions of probable synvolcanic age are rare and typically small, irregular bodies. The felsic volcanic rocks consist of coarse- and fine-grained pyroclastic rocks (2713±3 Ma; Krogh and Turek 1982) with lesser amounts of feldspar and quartz porphyritic flows (Ayres 1969b).

Lenticular iron-rich metasedimentary units (Ayres 1969b) occur throughout the assemblage and are associated with both the mafic and felsic volcanic rocks. Rocks in this assemblage are metamorphosed to greenschist grade and locally almandine amphibolite grade (Ayres 1969b). The Gamitagama Lake complex, a calc-alkalic stock which ranges in composition from olivine gabbro to syenite (Ayres 1969b; Choudhry 1981), intrudes the eastern half of the Gamitagama assemblage and imposes albite-epidote, hornblende and pyroxene hornfels metamorphism and cross folding (Ayres 1969b) on the surrounding volcanic rocks.

Conflicting facing data (e.g., pillow tops) indicates that the volcanic rocks may have undergone complex folding. Ayres (1969a) documented a series of major, tight to isoclinal, upright folds that plunge gently to moderately northwestward but whose mineral lineations plunge eastward at 65°, approximately orthogonal to these fold axes. Ayres (1969a) also noted that lineations steepened significantly in proximity to intrusions. These intrusions are also interpreted to have caused open to isoclinal cross folding of the early folds to generate a complex map pattern.

GARGANTUA ASSEMBLAGE

The Gargantua assemblage consists of: 1) sedimentary rocks occupying the southwestern part of the Gamitagama greenstone belt (*see* Figure 12.3); and 2) the Ryan Migmatite (metamorphic U-Pb age of 2615±9 Ma; Krogh and Turek 1982). The character of the boundary between the Gargantua assemblage and the Gamitagama assemblage to the north is unknown but is probably sheared.

The Gargantua assemblage is bounded on the south by granitoid rocks (Ayres 1969b) consisting of the Gargantua Harbour porphyritic granodiorite, equigranular tonalite and trondhjemite. There are several massive, equigranular to porphyritic quartz monzonite to granite intrusions within both the sedimentary assemblage and, locally, the Gamitagama volcanic assemblage to the north (*see* Figure 12.3).

The sedimentary rocks of the Gargantua assemblage are greywacke and siltstone with minor pebble and rare cobble conglomerate (Ayres 1969b). Conglomerate horizons are only found in close proximity to the contact with the Gamitagama assemblage volcanic rocks to the north. Ferruginous sedimentary rocks and lean iron formation are found intercalated with the greywacke and siltstone. Rocks of this assemblage are metamorphosed to greenschist grade and almandine amphibolite grade. The Ryan Migmatite is composed of approximately equal amounts of sedimentary and granitoid rocks (Ayres 1969b). The original distinction between the sediment-dominated Gargantua assemblage and the Ryan Migmatite is characterized by a southward gradation of sedimentary rocks into gneissic rocks, a corresponding southward increase in metamorphic grade and a greywacke-siltstone dominated (argillite-mudstone poor) sedimentary sequence, all features characteristic of metasedimentary subprovinces (e.g., Quetico Subprovince; *see* Williams, this volume). However, the scale of the Gargantua assemblage is much smaller than that of the Quetico Subprovince. The sedimentary rocks are comparable in size and type to those described from the Abitibi Subprovince in Quebec by Hubert (1990).

Facing data (e.g., graded bedding) suggest complex, tight to isoclinal folding within the sedimentary rocks. Ayres (1969a) documented an early, isoclinal fold generation that has a shallow to moderate northwest plunge. The style of folding within this assemblage is interpreted to be of similar character to that of the Gamitagama assemblage (Ayres 1969a).

MINERALIZATION

Mineralization in the Gamitagama greenstone belt consists of:

1. minor gold-bearing quartz veins associated with zones of high strain (e.g., Coutu occurrence; Ayres 1969b)
2. pyrrhotite-pentlandite-chalcopyrite associated with the Gamitagama Lake complex (e.g., Renner occurrence; Ayres 1969b)
3. chalcopyrite-bearing quartz veins associated with fault zones and sheared diabase dikes (e.g., Lost Gracie occurrence; Ayres 1969b)
4. lean, discontinuous, chert-magnetite iron formation

Michipicoten Greenstone Belt

The Michipicoten greenstone belt is the largest greenstone belt within the Wawa Subprovince and has been subjected to considerable field, geochemical and geochronologic study (e.g., Logan 1847; Goodwin 1962, 1963; Sage 1985, 1986). Economically, the belt has been one of the most significant; mining and exploration was directed towards gold and there is still an active siderite iron mine.

The Michipicoten greenstone belt is approximately 140 km long and reaches a maximum width of about 45 km. Supracrustal rocks within the belt strike in an easterly direction between the western extremity of the belt (Figure 12.4) and the Kapuskasing Structural Zone in the east (*see* Figure 12.2). Regional mapping delineated 3 discrete episodes of volcanism, originally called cycles (Sage 1986). The oldest episode (the Hawk assemblage—Cycle 1) is dated at approximately 2889 Ma (Turek et al. 1988), and is followed by 2 episodes of mafic to felsic volcanism (the Wawa assemblage—Cycle 2, and the Catfish assemblage—Cycle 3) which are dated at approximately 2750 and 2700 Ma, respectively. Original relations between these assemblages are commonly obscured by later shearing along their contacts. A description of the 3 assemblages is followed by remarks on their structure, alteration and metamorphism, and on the intrusive rocks within each assemblage.

HAWK ASSEMBLAGE

The Hawk assemblage (*see* Figure 12.4) is only known from 1 small, elongate area east of Wawa. Measuring only 5 km across and 25 km long, it is overlain to the north and west by rocks of the Wawa assemblage. To the south, across the Wawa–Hawk–Manitowik Lake fault, the Hawk assemblage is in contact with gneissic tonalite dated at 2747±7 Ma (Turek et al. 1982) and contains xenoliths of supracrustal rocks. The tonalitic gneisses are intruded by massive granodiorite and related phases of the Whitefish Lake batholith, which is dated at 2694±3 Ma (Turek et al. 1984).

The Hawk assemblage consists of pillowed and massive basaltic and peridotitic komatiites overlain by a calc-alkalic tuff, quartz-feldspar crystal tuff, lapilli tuff and breccia. The lower, mafic part of the episode was intruded by the Hawk granitic pluton, dated at approximately 2888±2 Ma (Turek et al. 1984), and by minor felsic sheets, dated at 2881 Ma (Turek et al. 1988).

The Hawk granitic pluton is, therefore, approximately the same age and perhaps comagmatic with the upper felsic part of volcanism. The volcanic rocks are capped by a thin chert-magnetite-sulphide iron formation and then by much younger volcanic rocks of the Wawa assemblage. The iron formation capping the Hawk assemblage may have precipitated from fluids associated with deep alteration of the volcanic pile during the terminal fumarolic stage of volcanism (Goodwin 1964; Lockwood 1986).

Komatiitic volcanic rocks of the 2.89 billion-year-old episode display a chondrite-normalized pattern permissive of pyroxene fractionation during magma evolution, and of garnet fractionation or a presence of garnet in the source material during the early stages of magma generation. The abundance of feldspar and quartz-feldspar crystal tuff in the felsic volcanic rocks indicates a fractionation of feldspar. This is supported by well-developed negative europium anomalies.

Table 12.2. Tectonic assemblages of the eastern and central Wawa Subprovince.

Assemblage; Probable Age; References	Contacts	Rock Types	Structure, Strain and Metamorphism	Mineralization
Dayohessarah–Kabinakagami; 2.75 to 2.70 Ga; Fenwick 1967; Siragusa 1972, 1985a, 1985b, 1977; Wilson 1989, 1990	all intrusive	volcanic rocks: pillowed, massive, and foliated mafic, rare felsic; sporadic ultramafic and gabbroic bodies; metasedimentary rocks: wacke predominant, minor iron formation and conglomerate	highly deformed rocks; variable, belt-parallel foliation trends; poorly constrained facing; upper greenschist to amphibolite facies	sulphide mineralization in mafic rocks; gold in belt-parallel Kabinakagami assemblage shear
Hemlo–Black River; 2.77 Ga; Milne 1968; Muir 1982a, 1982b, 1986, 1988, 1989; Muir and Elliott 1987; Siragusa 1984a, 1984b, 1985a, 1985b; Burk et al. 1986; Corfu and Muir 1989a; Pan 1990	south contact sheared against the Heron Bay assemblage; all others intrusive	lower volcanic rocks: pillowed, massive and foliated mafic; upper volcanic rocks: felsic to intermediate, with felsic centres; minor intrusions of gabbro and peridotite; metasedimentary rocks: wacke predominant, minor iron formation and conglomerate	well preserved to highly strained and schistose; variable foliation, orientations that parallel granitoid contacts; amphibolite facies	sulphide mineralization, gold and barite at or near mafic and/or felsic contacts, especially at strongly sheared southern boundary, e.g., Hemlo
Heron Bay; 2.70 Ga; Milne 1968; Muir 1982a, 1982b, 1986, 1988, 1989; Siragusa 1984a, 1984b, 1985a, 1985b; Patterson 1986; Pan and Fleet 1989; Corfu and Muir 1989a; Pan 1990	north contact sheared against Hemlo–Black River assemblage; all others intrusive	volcanic rocks: calc-alkalic, intermediate to felsic, thickens westward to volcanic centre at Heron Bay, iron-rich tholeiites of unknown relationship to other mafic rocks; minor intrusions of hornblendite and peridotite; metasedimentary rocks: wackes and shallow-water epiclastic sedimentary rocks predominant	generally well preserved; low strain, but strongly sheared; greenschist to amphibolite facies	stratiform units of molybdenite and gold-bearing, sheared, felsic and mafic volcanic rocks e.g., Peekongay property
Manitouwadge–Hornepayne; 2.75 to 2.70 Ga; Milne 1968, 1974; Giguere 1972; Bakker et al. 1985; Williams and Breaks 1989, 1990	all contacts are intrusive against the tonalite suite	lower volcanic rocks: dominantly mafic, including anorthosite, gabbro and ultramafic; upper volcanic rocks: sporadic intermediate to felsic tuff and agglomerate; metasedimentary rocks: wacke	highly strained schist and gneiss, especially at Quetico Subprovince boundary; early recumbent deformation during tonalite emplacement; later	massive sulphide deposits at sheared contact zone of mafic and felsic volcanic rocks at Manitouwadge

Table 12.2. Tectonic assemblages of the eastern and central Wawa Subprovince.

Assemblage; Probable Age; References	Contacts	Rock Types	Structure, Strain and Metamorphism	Mineralization
		predominant, oxide facies iron formation	deformation northeast-asymmetric folds and dextral shears; almandine amphibolite, granulite facies	
Schreiber; 2.75 to 2.70 Ga; Pye 1964; Walker 1967; Marmont 1983; Severin and Balint 1985; Schnieders 1987; Carter 1988a; Williams 1989, 1990	all contacts are intrusive except to the north; tectonic contacts against the Quetico Subprovince	lower volcanic rocks: diverse calc-alkalic agglomerate and tuff at central vent near Winston Lake; upper volcanic rocks: pillowed, massive and foliated tholeiitic, minor gabbroic and ultramafic intrusions; metasedimentary rocks: wacke predominant, minor shale, slate and conglomerate, sulphide iron formation at mafic rock-sedimentary rock contacts	generally well preserved; local east-trending stratigraphy and parallel shear zones; no consistent facing directions; structural repetition in the Jackfish area; high-strain zones along graphitic iron formation and felsic sheets; upper greenschist facies background; up to almandine amphibolite facies close to granitoid contacts	minor sulphide mineralization at granitoid contacts; massive zinc sulphide and associated alteration proximal to felsic centre at Winston Lake
Hawk; 2.89 Ga; Goodwin 1962; Sage 1985	overlain by the Wawa assemblage to the north; intrusive contacts to the south	volcanic rocks: pillowed, massive basaltic and peridotitic komatiite, calk-alkalic tuff and breccia; metasedimentary rocks: iron formation at the top	homoclinal, north-facing sequence; low strain	
Wawa; 2.75 Ga; Goodwin 1962, 1963; Sage 1985	unconformably overlies the Hawk assemblage; overlain unconformably by the Catfish assemblage; contacts are rarely original,and are usually sheared	lower volcanic rocks: pillowed, massive tholeiite and iron-rich tholeiite; upper volcanic rocks: calc-alkalic intermediate to felsic tuff, subvolcanic intrusion and breccia; capped by economically significant iron formation	well preserved, forms major fold structures, sheared assemblage contacts	banded iron formation of the McLeod Mine in the Helen Range; gold mineralization around the Jubilee stock

Table 12.2. Tectonic assemblages of the eastern and central Wawa Subprovince.

Assemblage; Probable Age; References	Contacts	Rock Types	Structure, Strain and Metamorphism	Mineralization
Catfish (Michipicoten and McCormick segments); 2.70 Ga; Goodwin 1962, 1963; Heather 1989; Heather and Arias 1987; Heather and Buck 1988; Sage 1985	unconformably overlies the Wawa assemblage; commonly sheared at contacts with the Wawa and Hawk assemblages; thrust-repeated as McCormick volcanic rocks; all other contacts are intrusive	volcanic rocks: massive and pillowed tholeiite and iron-rich tholeiite, overlain by iron formation and contemporaneous felsic to intermediate volcanic rocks; metasedimentary rocks: volcanic wackes	well preserved; forms major fold structures such as McCormick synform and Centre anticline	gold mineralization at Renabie and Goudreau–Lochalsh
Catfish (Miron segment); 2.70 Ga; Bennett and Thurston 1977; Reid et al., in press	contact against the Pukaskwa batholith to the north and Mishi assemblage to the south at the Mishibishu deformation zone	volcanic rocks: massive to foliated, rarely pillowed, iron-rich tholeiite with subordinate tholeiite and calc-alkalic rocks	primary structures not well preserved, north-dipping and north-facing homocline; amphibole lineations steeply plunging	gold-bearing quartz veins in high-strain zones that are related to the Mishibishu deformation zone; quartz-scheelite veins in enclaves of metasedimentary rock (Crane occurrence)
Catfish (Pukaskwa –Point Isacor segment); 2.70 Ga; Bennett and Thurston 1977; Reid et al., in press	sheared north contact with the Mishi assemblage, bounded to the south by the Floatingheart batholith (2693 Ma)	volcanic rocks: tholeiite and magnesium-rich tholeiite, calc-alkalic andesite to dacite	generally well preserved; low strain	lean iron formation; gold-bearing quartz veins in high-strain zones related to the Eagle River deformation zone; copper in faults and shear zones of Keweenawan age
Mishi; <2.70 Ga; Bennett and Thurston 1977; Reid et al., in press	sheared south contact with the Pukaskwa–Point Isacor segment; north contact against the Miron Lake segment	metasedimentary rocks: conglomerate, sandstone, wacke, argillite and mudstone	north-dipping, south-fining sequence; complexly folded, lineations plunge north	gold-bearing quartz veins in high-strain zones related to the Mishibishu deformation zone; Pb-Cu-Zn-Ag-bearing

Table 12.2. Tectonic assemblages of the eastern and central Wawa Subprovince.

Assemblage; Probable Age; References	Contacts	Rock Types	Structure, Strain and Metamorphism	Mineralization
	is the Mishibishu deformation zone			quartz veins related to post-Archean diabase dikes
Gamitagama; 2.70 – 2.72 Ga; Ayres 1969b; Choudry 1981; Krogh and Turek 1982	bounded to the south by the Gargantua assemblage and granites; bounded to the north by the Rabbit Blanket Lake intrusion (2668 Ma)	volcanic rocks: tholeiitic, basaltic to andesitic, lesser calc-alkalic rhyolitic to dacitic; gabbroic and dioritic sills	tight to isoclinal upright folds with gentle to moderate plunge; cross folding is related to intrusion	gold-bearing quartz veins in high-strain zones
Gargantua (including Ryan Migmatite); <2.70 Ga; Ayres 1969b; Choudry 1981; Krogh and Turek 1982	unknown boundary to the north; south boundary is against the Southern pluton	metasedimentary rocks: wacke, minor conglomerate, lean iron formation	complex, isoclinal folding, shallow to moderate northwest plunging axes; similar to that of the Gargantua assemblage	lean iron formation
Chapleau; <2.70 Ga; Percival 1990; Moser 1989; Leclair 1990; Thurston et al. 1977	conformable contacts paralleled by tonalite sheets	volcanic rocks: mafic rocks, now amphibolite and mafic granulite (volcanic origin?) metasedimentary rocks: quartzo-feldspathic gneiss and migmatite with much garnet and cordierite, rare conglomerate	layering transposed into regionally developed gneiss stratification, isoclinal folds and parallel layering, shallow to steep, northwesterly dips	minor sulphide

Table 12.3. Assemblages of the western part of the Wawa Subprovince.

Name	Rock Types, Primary Structures (described from base to top, where known)	Contacts	Interpreted Environment	Age (Ma), (if known)	References
Burchell assemblage	3 cycles of massive to pillowed basalt flows with local interbedded magnetite-quartz iron formation; tholeiitic pillowed flows predominant in the basal part of the lowermost cycle with some komatiitic flows; overlain by dacite to rhyolite tuff to pyroclastic breccia in separate, on-strike units; top of cycle 2 of massive to porphyritic rhyolite and rhyolitic pyroclastic breccia unit is up to 4 km wide, thins eastward to graded beds of distal felsic tuff, and traced over 75 km along strike; bedding tops from grading of tuff and pillow shapes.	locally identified fault contact with Quetico Subprovince metasedimentary rocks; tectonic back-to-back stratigraphic-facing relationship with the adjacent Greenwater assemblage forms the basis for assemblage division of Shebandowan greenstone belt	subaqueous to locally subaerial arc volcanism	>2733 (cycle 2)	Giblin 1964; Hodgkinson 1968; Morin 1973; Corfu and Stott 1986
Greenwater assemblage	3 cycles of massive to pillowed tholeiitic basalt and plagioclase-phyric basalt to andesite, and komatiitic flows; lowermost cycle capped by thin (200 m) dacitic and andesitic to rhyolitic tuff and lapilli tuff traced over 64 km along strike; magnetite-quartz or jasper iron formation provide bedding markers within felsic units of upper cycle 1 and lower basaltic part of cycle 2; iron formation in upper cycle 1 becomes amphibole-quartz-magnetite beds within distal andesitic tuff west of Greenwater Lake granodiorite pluton; cycle 2 capped by thick (7 km) sequence of dacite to rhyolite tuff, tuff breccia, and extrusive to subvolcanic quartz-feldspar porphyry; cycle 3 mainly tholeiitic flows; bedding tops mainly from pillow shapes and graded tuff beds	greater local folding of strata near contact with Burchell assemblage, south of Lower Shebandowan Lake; bedding-parallel fault contact with Burchell assemblage inferred but not observed	subaqueous to locally subaerial arc volcanism		Hodgkinson 1968; Carter 1990a, 1990b, 1990c; Chorlton 1987; Thurston 1984

Table 12.3. Assemblages of the western part of the Wawa Subprovince.

Name	Rock Types, Primary Structures (described from base to top, where known)	Contacts	Interpreted Environment	Age (Ma), (if known)	References
Saganagons assemblage	massive to pillowed basalt flows; minor rhyolite to dacite tuff breccia; local heterolithic breccia, clasts to 60 cm long	fault contact (Knife Lake fault) with Greenwater assemblage; locally tectonic, locally depositional contacts with adjacent Knife Lake assemblage	oceanic volcanism		Harris 1968
Shebandowan assemblage	folded metasedimentary rocks of conglomerates, arkose, mudstone, siltstone and wacke, with jasper-magnetite iron formation in the southern zone of 2 linear zones of this assemblage; calc-alkalic to alkalic (shoshonitic) flows and pyroclastic deposits as tuff, pyroclastic and debris flows; jasper-magnetite iron formation units form stratigraphic markers; bedding tops form graded sedimentary rocks and tuff beds, minor cross-beds	faulted, depositional contacts with underlying Burchell and Greenwater assemblages	alluvial-fluvial sedimentation and shallow subaqueous to subaerial continental volcanism in fault-bounded linear basins (grabens?)	~2689	Shegelski 1980; Corfu and Stott 1986; Carter 1990a, 1990b, 1990c
Knife Lake assemblage	folded lithic wacke turbidite, argillite and felsic tuff, and arkose and conglomerate; some basaltic flows interleaved with sedimentary rocks; resedimented conglomerate derived from erosion and unroofing of Saganaga pluton; wacke turbidite from mainly felsic volcanic source; graded and locally cross-bedded	proposed as tectonic equivalent to Shebandowan assemblage; faulted, depositional contacts upon Saganagons assemblage and Saganaga pluton; inferred tectonic fault contact with inferred extension of Greenwater assemblage into Minnesota (Newton Lake Formation)	submarine fan from eroded felsic volcanic and late plutonic source		Green 1970; Ojakangas 1972a, 1972b, 1979; Sims 1985

Chondrite-normalized rare earth element (REE) plots for the Hawk granitic pluton, dated at 2888 Ma, are similar to those obtained for the Hawk felsic volcanic rocks. This supports the interpretation that the intrusive and volcanic rocks are comagmatic.

$^{207}Pb/^{204}Pb$ ratios determined for mineralized veins within the Hawk volcanic rocks are high for their age, implying the presence of an earlier crustal component (Thorpe 1987). Rocks that could represent this earlier crust have not been identified in the Wawa region.

WAWA ASSEMBLAGE

This extensive assemblage structurally underlies much of the Catfish assemblage and overlies the Hawk assemblage in the south (*see* Figure 12.4). Commonly, the original stratigraphic relations between the Catfish and the Wawa assemblages are obscured by shearing along their contact. The Wawa assemblage lies in apparent conformable contact with the Hawk assemblage along Loonskin Lake (*see* Figure 12.4). The metavolcanic rocks along this contact are

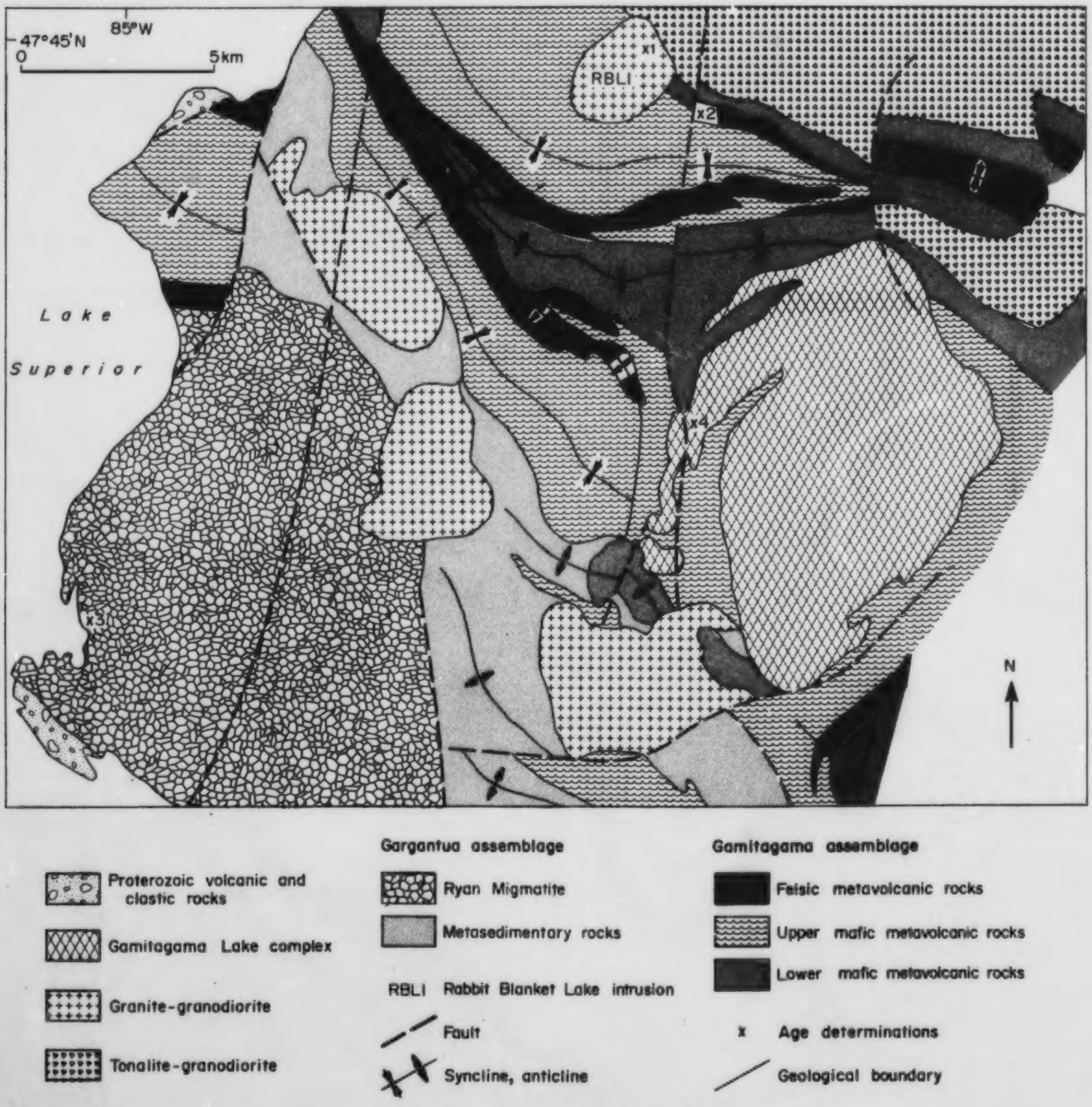

47°45'N
85°W
0
5km
RBLI
x1
x2
x3
x4
Lake
Superior
N
Proterozoic volcanic and clastic rocks
Gamitagama Lake complex
Granite-granodiorite
Tonalite-granodiorite
Gargantua assemblage
Ryan Migmatite
Metasedimentary rocks
RBLI Rabbit Blanket Lake intrusion
Fault
Syncline, anticline
Gamitagama assemblage
Felsic metavolcanic rocks
Upper mafic metavolcanic rocks
Lower mafic metavolcanic rocks
x Age determinations
Geological boundary

Figure 12.3. Geological map showing tectonic assemblages of the Gamitagama greenstone belt (*modified from* Ayres 1969b).

strongly sheared and the contact is likely tectonic rather than depositional.

The lower part of the Wawa assemblage consists of massive and pillowed magnesium- and iron-rich tholeiitic volcanic rocks. South of Wawa, these intermediate to mafic volcanic rocks are separated from overlying calc-alkalic intermediate to felsic volcanic rocks by polymictic breccia and epiclastic tuff of intermediate composition. The polymictic breccia probably represents lahars or mass flow deposits marking the onset of felsic calc-alkaline volcanism. Thin units of chert, pyrite and graphite may be associated with these rare clastic rocks.

The intermediate to felsic volcanic rocks that form the upper part of the Wawa assemblage consist of tuff, quartz-feldspar crystal tuff, lapilli tuff, oligomictic and polymictic breccia and scarce spherulitic flows. The stratigraphic thickness of intermediate to felsic volcanic rocks varies considerably. At Wawa, a 2 km thickness of intermediate to felsic volcanic rocks diminishes to 300 m over a distance of 10 km eastward. Stratigraphic thinning and fining eastward, from the thick stratigraphic section near Wawa, is suggestive of a more proximal vent facies there.

The Jubilee stock (*see* Figure 12.4) is located at the base of this thick section of intermediate to felsic volcanic rocks. The stock contains xenolith blocks and large irregular masses of volcanic rock indicating final emplacement at a high structural level.Therefore, the stock is considered to be the subvolcanic equivalent of intermediate to felsic volcanic rocks of the Wawa assemblage . The Jubilee stock has been dated at approximately 2745±3 Ma (Sullivan et al. 1985), not dissimilar to the ages of 2744 and 2749 Ma obtained on the intermediate to felsic volcanic rocks of the Wawa assemblage (Turek et al. 1982; Corfu and Sage 1987).

Capping the Wawa assemblage volcanic rocks is a 100 to 150 m thick section of iron formation with a total strike length exceeding 100 km. The formation of the iron formation is a result of both distal clastic sedimentation and chemical precipitation. The occurrence of iron formation overlying recently formed felsic volcanic rocks may be considered to be the result of chemical precipitation on a volcanic platform from fluids that produced hydrothermal alteration at depth. This iron formation accounts for all commercially exploitable iron deposits within the Michipicoten greenstone belt and is renowned for its well-developed ore-grade carbonate and sulphide facies deposits.

Production in the Wawa camp has continued from 1900 to the present. The major producing mine, the McCleod Mine, has produced over 79 million tons of iron from 1938 through 1985 (Berdusco *in* Gross and Donaldson 1990). The iron formation consists of at least 5 facies whose distribution is consistent throughout the greenstone belt, even though the degree of development of each facies varies radically from site to site. From the stratigraphic base to top, the facies are massive siderite grading upward to massive pyrite and pyrrhotite which grades upward into bedded chert-magnetite-wacke. The chert-magnetite-wacke grades upward to chert-wacke which grades upward into siliceous, graphitic, pyritic argillite.

The presence of abundant graphite in the upper part of the iron formation stimulated geochemical and carbon-sulphur isotopic studies, which concluded that biological activity played a role in the genesis of the iron formation (Goodwin et al. 1976; Karkhanis et al. 1980; Thode and Goodwin 1983; Goodwin et al. 1985). A study of oxygen isotopes by LeSeelleur (1989) showed that iron formation deposition took place at temperatures of less than 200°C. The iron and minor manganese content in the carbonate facies iron formation is indicative of a low-temperature, distal deposit when compared to a mid-oceanic ridge fumarolic environment. The position of the iron formation between 2 volcanic cycles suggests an intimate relationship between volcanism and the possible volcanic-fumarolic activity related to its formation.

Using amygdule size and volume, a shallowing of the depositional basin with time is evident in the mafic volcanic rocks within the Wawa assemblage. Mafic flows contain few amygdules at the base of the sequence, but amygdules are numerous and have a size of 1 cm or more at the top of the sequence. The presence of such large and abundant amygdules may indicate a low pressure, shallow water environment (Moore 1965). In support of the theory of a shallow-water environment, the supracrustal rocks contain massive siderite in which stromatolitic structures have been replaced by pyrite (Hoffman et al., in press).

The magnesium- and iron-rich tholeiitic rocks of the Wawa assemblage, dated at 2750 Ma, display chondrite-normalized REE patterns consistent with amphibole fractionation and are similar to those obtained from rocks in an island-arc or continental-arc setting (Sylvester et al. 1987). The abundance of feldspar and quartz-feldspar crystal tuff indicates feldspar fractionation in the felsic melts. This is supported by well-developed negative europium anomalies, suggestive of feldspar fractionation during felsic magma evolution.

CATFISH ASSEMBLAGE (MICHIPICOTEN AND McCORMICK SEGMENTS)

The Catfish assemblage is by far the most prominent assemblage within the Michipicoten and neighbouring greenstone belts (*see* Figure 12.4). The Catfish assemblage structurally overlies the Wawa assemblage and, due to thrusting, the stratigraphy has been locally repeated. This feature is now preserved on the east limb of the McCormick synform and is known as the McCormick Lake volcanic rocks.

Rarely (e.g., Sir James Pit, Wawa) does the Catfish assemblage show an unconformable relationship preserved against the underlying Wawa assemblage (*see* Figure 12.4; Arias and Helmstaedt 1990). Commonly, this relationship is obscured by shearing. Some of the eastern part of the Michipicoten greenstone belt consists of the Catfish assemblage, forming highly metamorphosed septa and enclaves within foliated to gneissic tonalite and granodiorite (*see* Figure 12.4). The subdivision of this region into the Wawa and Catfish assemblage is confounded by the lack of sufficiently detailed mapping.

Lying above the Wawa assemblage, the lower part of the Catfish assemblage is composed of massive and pillowed magnesium- and iron-rich tholeiitic flows (*see* Figure 12.4). These mafic rocks are indistinguishable from the mafic rocks that form the lower part of the Wawa assemblage volcanic rocks and have only been recognized as younger by a combination of structural and geochronologic studies. Lying stratigraphically above the mafic volcanic rocks are sedimentary rocks and intermediate to felsic volcanic rocks in conformable contact. The intermediate to felsic volcanic rocks consist of tuff, lapilli tuff, coarse to very coarse breccia and abundant quartz-feldspar crystal

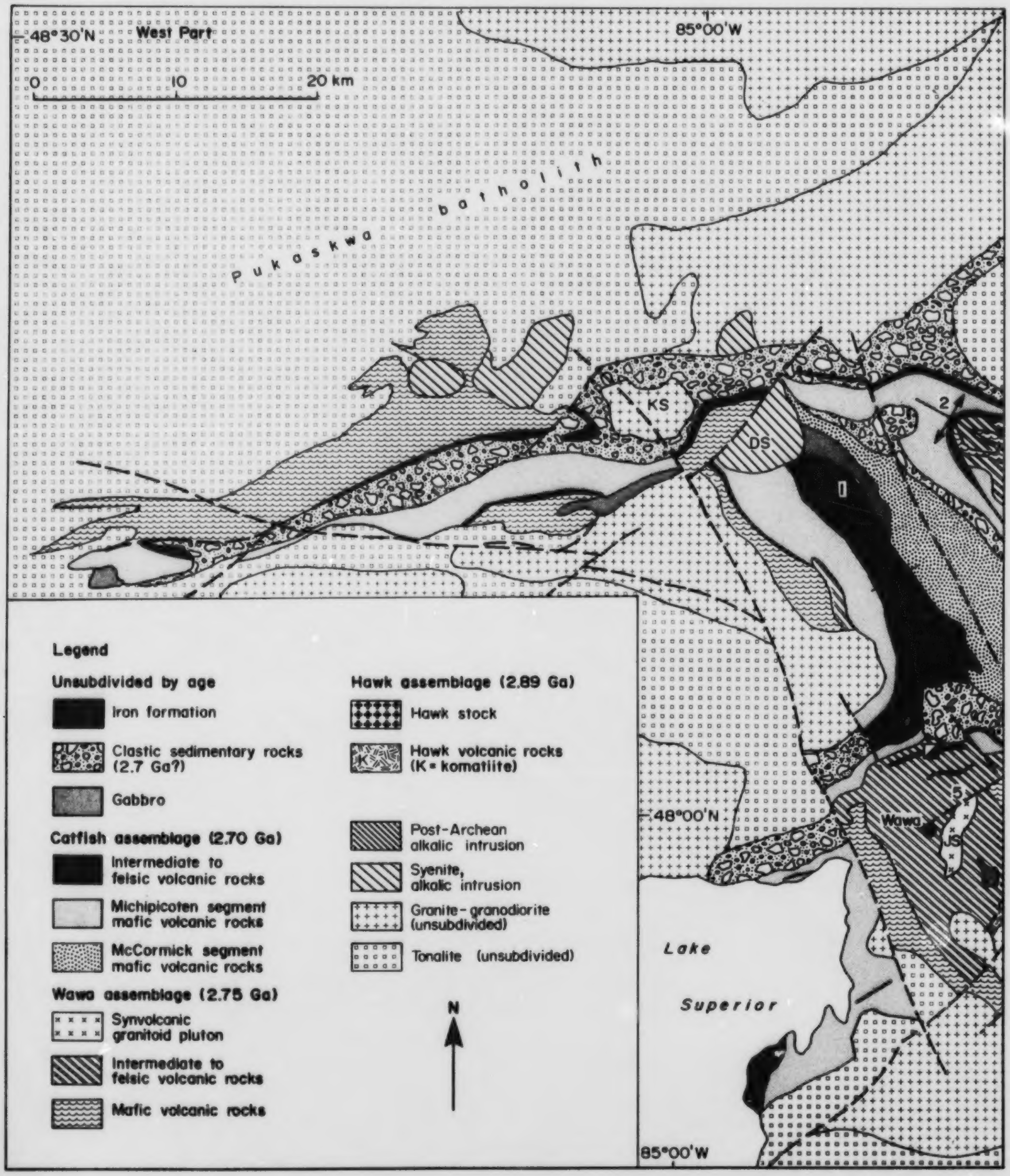

48°30'N
West Part
0
10
20 km
85°00'W
Pukaskwa batholith
KS
DS
2
Wawa
JS
48°00'N
Lake Superior
85°00'W
N
Legend
Unsubdivided by age
Iron formation
Clastic sedimentary rocks (2.7 Ga?)
Gabbro
Catfish assemblage (2.70 Ga)
Intermediate to felsic volcanic rocks
Michipicoten segment mafic volcanic rocks
McCormick segment mafic volcanic rocks
Wawa assemblage (2.75 Ga)
Synvolcanic granitoid pluton
Intermediate to felsic volcanic rocks
Mafic volcanic rocks
Hawk assemblage (2.89 Ga)
Hawk stock
Hawk volcanic rocks (K = komatiite)
Post-Archean alkalic intrusion
Syenite, alkalic intrusion
Granite-granodiorite (unsubdivided)
Tonalite (unsubdivided)

Figure 12.4. Geological map showing the distribution of major rock types and the tectonic assemblages of the Michipicoten greenstone belt (*see* Figure 12.2). Note the location of the cross section A-A′, depicted in Figure 12.5.

tuff. The tuff and crystal tuff are the dominant sediment source for the contemporaneous, interdigitating sedimentary units.

Tholeiites of the 2700 million-year-old Catfish assemblage (Turek et al. 1984) display chondrite-normalized REE patterns consistent with amphibole fractionation. Associated felsic volcanic rocks contain abundant feldspar and quartz-feldspar crystal tuff, supportive of the fractionation of feldspar in the felsic melts. The felsic rocks display well-developed negative europium anomalies, suggesting

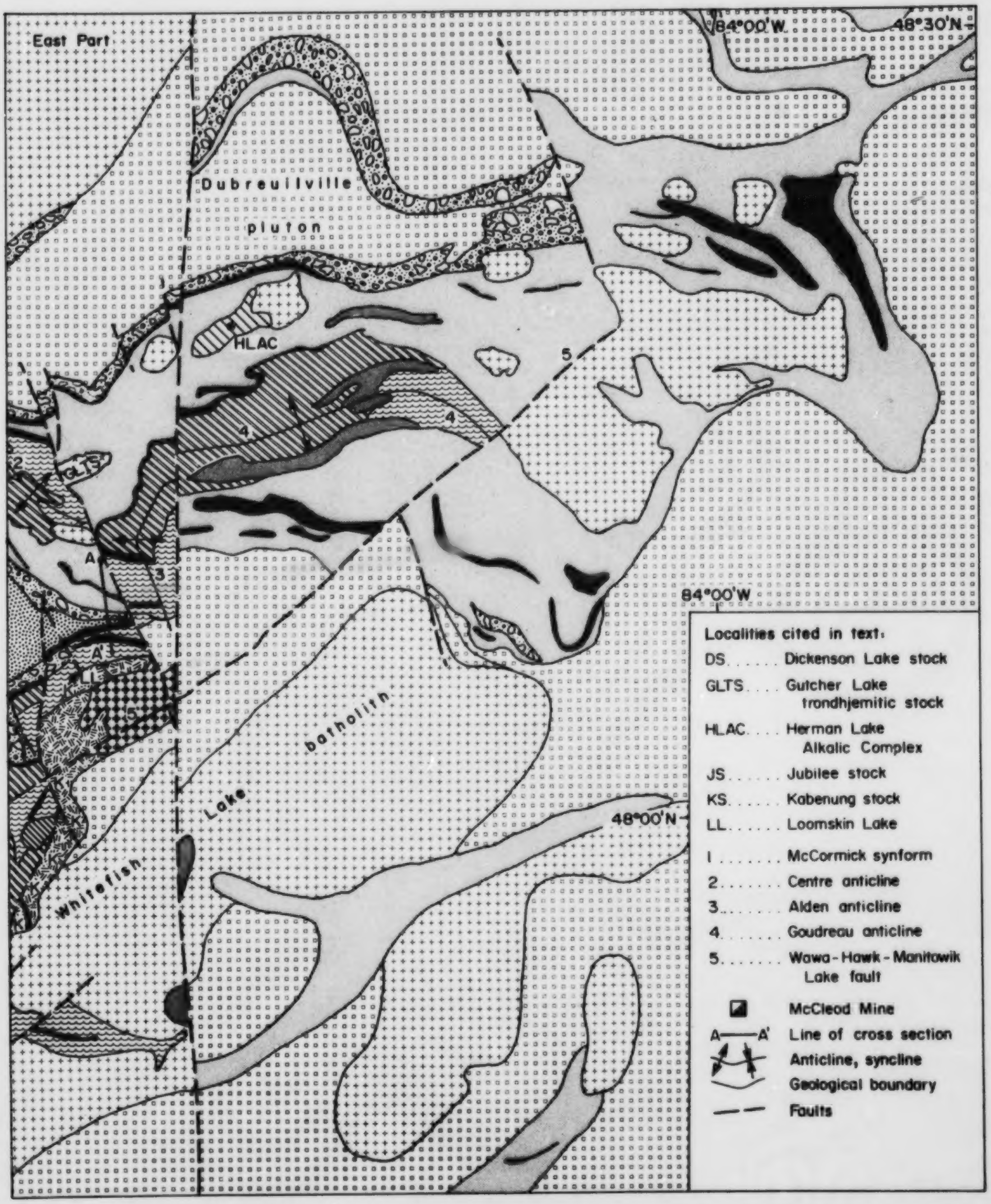

East Part
84°00'W
48°30'N
Dubreuilville
pluton
HLAC
GLTS
A
A'
LL
1
2
3
4
5
Whitefish
Lake
batholith
48°00'N
84°00'W
Localities cited in text:
DS....... Dickenson Lake stock
GLTS.... Gutcher Lake trondhjemitic stock
HLAC.... Herman Lake Alkalic Complex
JS....... Jubilee stock
KS....... Kabenung stock
LL....... Loomskin Lake
1........ McCormick synform
2........ Centre anticline
3........ Alden anticline
4........ Goudreau anticline
5........ Wawa-Hawk-Manitowik Lake fault
McCleod Mine
A——A' Line of cross section
Anticline, syncline
Geological boundary
Faults

Figure 12.4. Geological map showing the distribution of major rock types and the tectonic assemblages of the Michipicoten greenstone belt (*see* Figure 12.2). Note the location of the cross section A-A', depicted in Figure 12.5.

feldspar fractionation played a major role in felsic magma evolution. Chondrite-normalized REE data obtained on these volcanic rocks are similar to patterns obtained on rocks in an arc or continental-arc setting (Sylvester et al. 1987).

Sedimentary rocks overlying the Catfish assemblage volcanic rocks consist of argillite, siltstone, wacke, subarkose and conglomerate. Their contact with the volcanic rocks of the Catfish assemblage is a gradational depositional contact. The unit has been known as the Doré conglomerates (Logan 1863). The sedimentary rocks are well bedded with the lower parts of the sedimentary stratigraphy dominantly consisting of siltstone and wacke that display mudchip breccias, soft-sediment slump structures, excellent flame structures and grain-graded beds. The upper part of the sedimentary section is dominated by wacke and subarkose which may display cross-bedding, rare ripple marks and even rarer desiccation cracks.

The presence of desiccation cracks indicates a subaerial environment and since the sedimentary rocks are contemporaneous with 2700 million-year-old volcanism, volcanism is also likely to be subaerial for there is a dominance of fragmental rocks suggestive of a Plinian style of eruption (Fisher and Schmincke 1984). The sedimentary rock section contains deep-water turbidite deposits succeeded by shallow-water braided stream deposits (Neale 1981; Thomas 1984). All sedimentary rocks are immature and contain lithic fragments, abundant feldspar and display angular to subrounded grain shapes. Conglomerate associated with the sedimentary rocks increases dramatically in thickness westwards, and contains buff, medium- to coarse-grained trondhjemite cobbles and boulders which also increase in number and size towards Lake Superior. These granitic boulders were dated (Corfu and Sage 1987) at 2698±2 Ma, indistinguishable from ages of 2696±2 Ma and 2698±11 Ma (Turek et al. 1982, 1984) on the intermediate to felsic volcanic rocks of the Catfish assemblage. The trondhjemitic clasts likely represent eroded detritus from the intrusive equivalent of the intermediate to felsic volcanic rocks of the Catfish assemblage not presently exposed but lying beneath younger Proterozoic rocks and Lake Superior.

PLUTONIC ROCKS

Some plutonic rocks within the greenstone belt have been interpreted to be the subvolcanic equivalents of the overlying volcanic rocks (*see* Figure 12.4), whereas other plutonic rocks clearly intruded after a period of sedimentation and deformation. For example, the granitic Hawk stock is the postulated source for the Hawk assemblage; similarly, the Jubilee stock underlies and intrudes the felsic volcanic rocks of the Wawa assemblage. The REE content of the Jubilee stock is enriched to 10 times that of the older Hawk stock. These data support the possibility that the Jubilee stock had a crustal source, whereas the Hawk stock had a mantle or amphibolitic source. Chondrite-normalized REE patterns of younger stocks display similar REE contents as the Jubilee stock but lack a negative europium anomaly that characterized the older volcanic-equivalent stocks. These younger stocks may, therefore, represent unfractionated crustal melts.

Along the northern part of the Michipicoten greenstone belt, numerous stocks, ranging from trondhjemite to nepheline-cancrinite syenite, intrude supracrustal rocks. One of the oldest stocks, at 2722±1 Ma (Turek, Sage and Van Schmus 1990), is the Gutcher Lake trondhjemitic stock, which was emplaced between the formation of the Wawa and Catfish assemblages. Most of these intrusions display massive, equigranular to porphyritic textures and range in age from approximately 2670 to 2700 Ma. Many of these postfolding stocks lack a negative europium anomaly and show no evidence of fractionation. In contrast, plutons that are coeval with the Wawa and Hawk volcanism (Jubilee and Hawk stocks) do display a negative europium anomaly (along the southern side of the Michipicoten greenstone belt) thus indicating feldspar fractionation.

ALTERATION

Alteration of supracrustal rocks within the Michipicoten greenstone belt is pervasive and can be subdivided into that occurring concurrent with volcanism and that accompanying regional metamorphism. Volcanic rocks in the Michipicoten greenstone belt exhibit extensive zones containing chloritoid that are equally well developed and compositionally similar in both the mafic and felsic extrusive rocks but only rarely seen within mafic intrusions. The main development of chloritoid straddles the iron formation and caps the Wawa assemblage. Chloritoid is, however, known to exist from as low in the stratigraphic section as the upper part of the Wawa assemblage to as high as felsic volcanism within the overlying Catfish assemblage.

Rocks that contain chloritoid porphyroblasts are commonly heavily carbonatized, but not all heavily carbonatized rocks contain chloritoid. The chloritoid is common in proximity to the iron formation, but there are numerous occurrences of chloritoid-bearing rock without associated iron formation. Chloritoid-bearing zones are conformable to major lithologic units several kilometres in strike length. The zones lack evidence of regional structural control and may contain many different flow units displaying a variety of primary textures. Lockwood (1986) concluded that the alteration was of low temperature (less than 325°C), required the presence of carbon dioxide, involved low water to rock ratios, represented low-velocity, unfocussed discharge and developed syntectonically to posttectonically in response to regional metamorphism of previously altered volcanic rocks. McMillan (1981) suggested that nontronite was the precursor to chloritoid, but Lockwood (1983) preferred smectite or iron chlorite.

Tourmaline has been locally introduced into massive and pillowed mafic volcanic rocks accompanied by geochemically anomalous gold (Teal 1986).

In one small area below the iron formation that caps the felsic volcanic rocks within the Wawa assemblage, alteration involved the addition of CaO, K_2O, Al_2O_3 and strontium, and complementary removal of SiO_2, Na_2O, Fe_2O_3 and CO_2 (Lisowyk 1987).

The central part of the Michipicoten greenstone belt has been extensively carbonatized; it affects all rock types and may be so pervasive locally that it makes the interpretation of an original rock type difficult.

STRUCTURAL DEVELOPMENT

Goodwin (1964) prepared the first structural synthesis of the Michipicoten greenstone belt, recognizing that the supracrustal rocks were folded around east-trending axes of the Centre and Goudreau anticlines (*see* Figure 12.4) and proposed that they were subsequently refolded about northwest-trending axes. The supracrustal rocks have been cut by numerous, late, sinistral wrench faults which display a decreasing amount of stratigraphic offset from south to north. These north-striking transverse faults are commonly diabase filled. The area has remained tectonically stable, apart from diabase emplacement in the Late Archean and during the Proterozoic.

Percival and Card (1983, 1985) have suggested that the Kapuskasing Structural Zone, lying at the eastern border of the Wawa Subprovince, is a tilted block exposing the upper 20 km of crust in which deeper structural levels below the Michipicoten greenstone belt are successively exposed to the east.

Regional mapping has recently modified the interpretation of Goodwin (1964); detailed structural studies now emphasize the sheared or modified unconformable boundary relations between and within assemblages (McGill and Shrady 1986; Arias and Heather 1987; Heather and Arias 1987; Heather and Buck 1988; Arias and Helmstaedt 1990). Structural interpretations currently include recognition of a modified fold-thrust belt geometry for rocks comprising the Wawa and Catfish assemblages to the northeast of Wawa (Figure 12.5; Arias and Helmstaedt 1990).

Numerous dip-slip and strike-slip faults and shear zones, a number of which are mineralized with gold (Arias and Heather 1987; Heather and Arias 1987; Heather and Buck 1988), are subparallel to stratigraphic units or define assemblage boundaries (*see* Figure 12.4). The faults and shear zones separate assemblages with similar, or contrasting, stratigraphic facing and structural style. The shear zones may be several hundred metres wide, with deformation taking place at formational contacts or along bedding surfaces. Movement senses and amounts are often indeterminate and in some cases multiphase. Much of this shearing took place early in the deformational history and was affected by later, folding deformations.

In the area northeast of Wawa, McGill and Shrady (1986) recognized bedding-plane parallel thrust faults and proposed that significant early horizontal transport had taken place within the original north-facing homoclinal supracrustal assemblage. Arias and Helmstaedt (1990) recorded not only original unconformable relationships between the Wawa and Catfish assemblages but have recognized a regionally inverted stratigraphic sequence representing the lower limb of a recumbent fold. Subsequent regional shortening of this sequence imposed easterly trending folds and high-angle reverse faults upon it (*see* Figure 12.5).

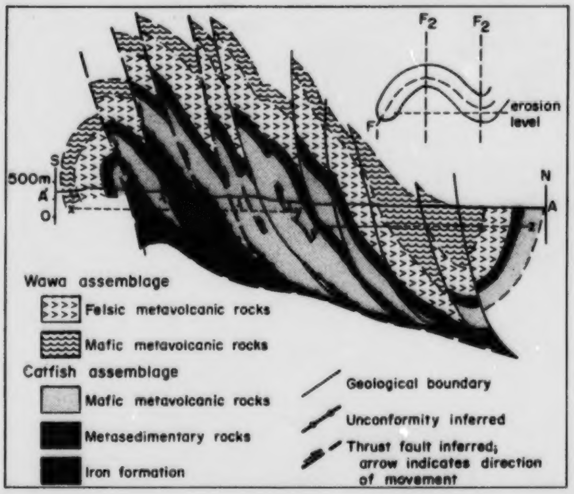


Figure 12.5. Composite structural cross section through the central part of the Michipicoten greenstone belt (*modified from* Arias and Helmstaedt 1990). Section x-y is an admissible section based on Arias and Helmstaedt (1990). Section y-z is a schematic section based on compilation (Goodwin 1963; Sage 1985). The sketch at lower left explains the present configuration of the belt as a regional scale recumbent fold (F_1) refolded about upright F_2 folds. Imbricate thrusts are considered to be contemporaneous with F_2 folding.

Precise zircon geochronology (Turek, Keller and Van Schmus 1990) has shown that volcanic rocks on the east limb of the McCormick synform (*see* Figure 12.4) represent a thrust-repeated unit of the Catfish assemblage and are here defined as the McCormick segment of that assemblage.

Structural data, though incomplete, is supportive of an initial period of recumbent folding and thrusting along or parallel with lithologic boundaries. This was followed by upright folding and high-angle reverse faulting, followed by folding, resulting, in part, from granitoid emplacement.

Mishibishu Greenstone Belt

The Mishibishu greenstone belt (Evans 1942; Bennett and Thurston 1977; Bowen and Logothetis 1985; Bowen 1986; Reid and Reilly 1987; Reid et al., in press) is a broad, arcuate belt approximately 60 km in length and 18 km wide (*see* Figure 12.2). The belt can be subdivided into 3 lithotectonic units; they are the Miron segment and the Pukaskwa–Point Isacor segment of the Catfish assemblage and the Mishi assemblage (*see* Figures 12.2 and 12.6). The 2 volcanic segments (Miron and Pukaskwa–Point Isacor) are lithologically and geochronologically similar to, and therefore probably correlative with, the Catfish assemblage of the Michipicoten greenstone belt and have been defined as such on the accompanying Tectonic Assemblages of Ontario map (*see* Map 2577, map case).

CATFISH ASSEMBLAGE

Miron Segment

The Catfish assemblage, Miron segment, occupies the northern flank of the Mishibishu Lake greenstone belt and is thought to include both the Bonamie Cove and Marjorie enclaves (*see* Figure 12.6). The segment is bounded to the north by tonalite (2721±4 Ma; Turek, Keller and Van Schmus 1990), granodiorite, quartz diorite and local hornblende granite and quartz monzonite of the Pukaskwa batholith (Reid et al., in press).

The contact between the Pukaskwa batholith and the supracrustal rocks of the Miron segment is a complex zone of granitoid sheeting and enclaves of migmatized supracrustal rocks. The Miron segment is bounded to the south by extensive sedimentary rocks of the Mishi assemblage (*see* Figure 12.6). The boundary between the Miron segment and the Mishi assemblage is marked by the Mishibishu deformation zone (MDZ), a regionally extensive zone of high strain (*see* Figure 12.6).

The Miron segment is dominated by iron-rich tholeiites intercalated with minor tholeiitic and calc-alkalic basalts and andesites. There is a distinct lack of dacitic and rhyolitic volcanic rocks within the Miron segment. These mafic volcanic rocks are generally massive to foliated and commonly feldspar porphyritic, with pillows being extremely rare. Medium- to coarse-grained, synvolcanic mafic sills are common and difficult to distinguish from coarse-grained flows. Mafic to intermediate pyroclastic rocks, including crystal tuff, lapilli tuff and tuff breccia, are locally intercalated with flows. The Bonamie Cove and Marjorie enclaves contain amphibolitized mafic volcanic rocks and minor sedimentary rocks displaying varying degrees of migmatization. Rocks in this segment are metamorphosed to amphibolite grade within 1 km of the Pukaskwa batholith with the remainder being of greenschist grade.

Facing data, such as pillow tops, indicate that the Miron segment is a north-dipping, north-facing sequence of rocks. The regional foliation and the lithologic layering are subparallel and mimic the shape of the greenstone belt. Dimensional lineations plunge 50° to 80° to the north, but there have been no folds outlined within this segment (Reid et al., in press).

Pukaskwa–Point Isacor Segment

The Pukaskwa–Point Isacor segment occupies the southern part of the Mishibishu greenstone belt (*see* Figure 12.6). This complex volcanic segment stretches from Pukaskwa Depot in the west to Point Isacor in the east (*see* Figure 12.6). The contact between the sedimentary rocks of the Mishi assemblage and rocks of the Pukaskwa–Point Isacor segment

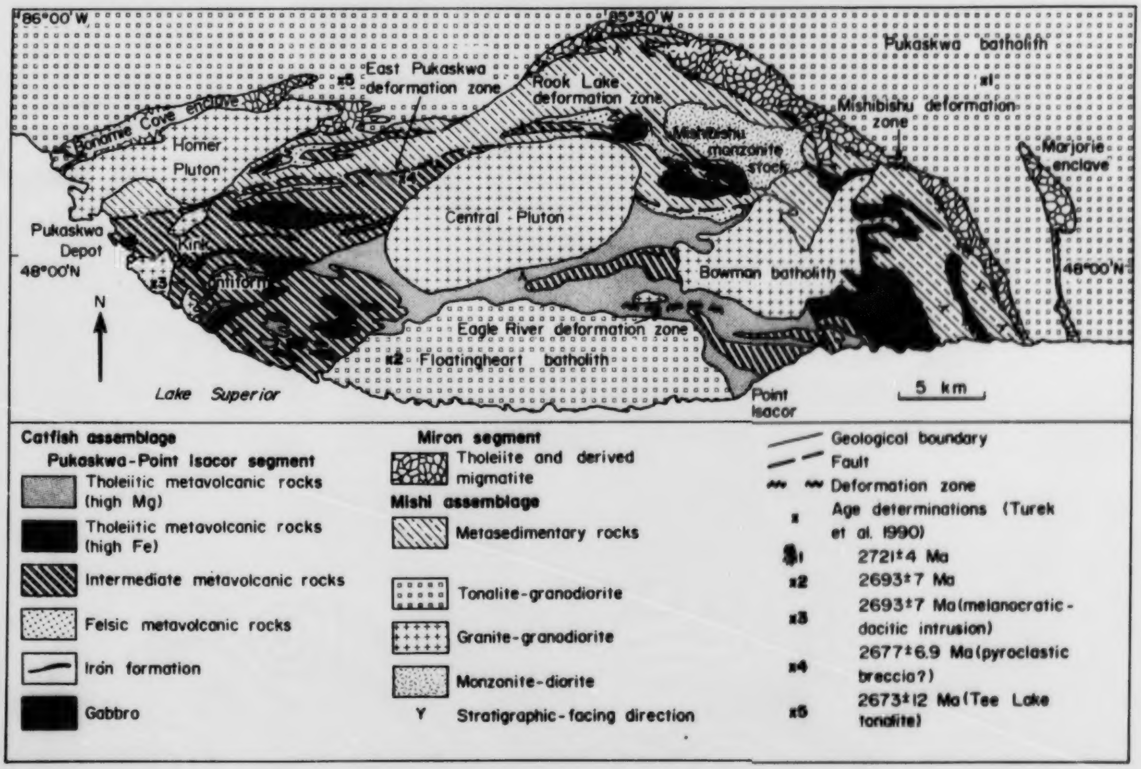

86°00'W
85°30'W
Pukaskwa batholith
x1
x5
East Pukaskwa deformation zone
Rook Lake deformation zone
Mishibishu deformation zone
Bonamie Cove enclave
Homer Pluton
Mishibishu monzonite stock
Marjorie enclave
x4
Central Pluton
Pukaskwa Depot
48°00'N
Kink
x3
Antiform
Bowman batholith
48°00'N
N
Eagle River deformation zone
x2 Floatingheart batholith
Lake Superior
Point Isacor
5 km
Catfish assemblage
Pukaskwa-Point Isacor segment
Tholeiitic metavolcanic rocks (high Mg)
Tholeiitic metavolcanic rocks (high Fe)
Intermediate metavolcanic rocks
Felsic metavolcanic rocks
Iron formation
Gabbro
Miron segment
Tholeiite and derived migmatite
Mishi assemblage
Metasedimentary rocks
Tonalite-granodiorite
Granite-granodiorite
Monzonite-diorite
Y Stratigraphic-facing direction
Geological boundary
Fault
Deformation zone
x Age determinations (Turek et al. 1990)
x1 2721±4 Ma
x2 2693±7 Ma
x3 2693±7 Ma (melanocratic-dacitic intrusion)
x4 2677±6.9 Ma (pyroclastic breccia?)
x5 2673±12 Ma (Tee Lake tonalite)

Figure 12.6. Geological map showing tectonic assemblages of the Mishibishu greenstone belt (*modified from* Reid et al., in press). The locations of the East Pukaskwa, Eagle River, Mishibishu and Rook Lake deformation zones are cited in the text.

appears locally to be a system of faults and ductile shear zones. The Pukaskwa–Point Isacor segment is bounded to the south (*see* Figure 12.6) by both Lake Superior and foliated to massive, biotite ± hornblende tonalite of the Floatingheart (Southern) batholith (2693±7 Ma; Turek, Keller and Van Schmus 1990). The structural style of this segment is characterized by regional folding and the local development of zones of high strain (e.g., Eagle River deformation zone, ERDZ; *see also* Figure 12.6). In general, the rocks within this segment exhibit less strain than the majority of the rocks within the previously discussed assemblages.

The Pukaskwa–Point Isacor segment consists of magnesium-rich tholeiite, calc-alkalic andesite, dacite and minor rhyolite, minor sedimentary rocks and lean chert-magnetite iron formation (*see* Figure 12.6). Magnesium-rich tholeiites predominate in the eastern two-thirds of the segment, whereas calc-alkalic flows, porphyritic flows, crystal tuffs and coarse pyroclastic rocks, of andesitic to dacitic composition, occupy much of the western one-third of the segment.

Reid et al. (in press) suggested that 3 melanocratic quartz-feldspar intrusions of dacitic composition (2693±7 Ma; Turek, Keller and Van Schmus 1990), located near Pukaskwa Depot (*see* Figure 12.6), are volcanic centres for the large volume of calc-alkalic volcanic rocks. Gabbroic to dioritic intrusions are common in the western one-third of the Pukaskwa–Point Isacor segment. The calc-alkalic volcanic rocks in the western part of the segment are complexly intercalated with the magnesium-rich tholeiites in the east. Rhyolitic volcanic rocks are rare, forming a discontinuous "horizon" of small, isolated, elongate bodies that occur close to the Mishi assemblage to the north (*see* Figure 12.6).

Regionally extensive, lean, chert-magnetite iron formation, like the felsic volcanic rocks, are confined to the northern half of the Pukaskwa–Point Isacor segment. Sedimentary rocks are rare and are of limited areal extent. Rocks of the Pukaskwa–Point Isacor segment are metamorphosed to greenschist metamorphic grade, while amphibolite grade is attained within the contact metamorphic aureoles adjacent to the Central pluton and Floatingheart (Southern) batholith.

The Pukaskwa–Point Isacor segment of supracrustal rocks contains regional scale folds (Reid et al., in press), the largest being the Kink antiform (*see* Figure 12.6). Where relatively unaffected by these folds, the volcanic rocks form a north-dipping, north-facing sequence. Located at the western end of the belt, the Kink antiform has a shallow to moderate, north plunge. Farther to the east, several folds have east-trending axial traces and moderate to steep (i.e., 45° to 70°) east-plunging fold axes.

The dominant schistosity is axial planar to these east-striking folds, and the schistosity is folded by the Kink antiform, indicating 2 stages of folding. Dimensional lineations on this schistosity range from shallow north plunging in the west, to moderate east plunging in the east.

MISHI ASSEMBLAGE

The Mishi assemblage consists of sedimentary rocks separating the Miron segment from the Pukaskwa–Point Isacor segment (*see* Figure 12.6). The boundary between the Mishi assemblage and the Miron segment of the Catfish assemblage is marked by the regionally extensive Mishibishu deformation zone (MDZ), and is cut by the Homer pluton, which consists of granite, quartz monzonite and tonalite intrusions (2673±12 Ma; Turek, Keller and Van Schmus 1990). These intrusions also separate the Bonamie Cove enclave from the main Mishibishu greenstone belt (*see* Figure 12.6).

The boundary between the Mishi assemblage and the Pukaskwa–Point Isacor segment of the Catfish assemblage is marked by a system of faults and deformation zones (*see* Figure 12.6). The northern parts of both the Central pluton and the Bowman batholith intrude sedimentary rocks of the Mishi assemblage, whereas the Mishibishu monzonite stock is confined within the sedimentary rocks (*see* Figure 12.6).

The Mishi assemblage is a north-dipping, southward-fining sequence of metasedimentary rocks that exhibit a complex pattern of facing directions based on graded bedding.

The sedimentary rocks of the Mishi assemblage consist of intercalated polymictic conglomerate, oligomictic quartz granule conglomerate, sandstone, wacke, mudstone and argillite. In general, coarser clastic rocks predominate within the northern part of the Mishi assemblage, while the southern part is dominated by finer-grained sedimentary rocks. The polymictic conglomerates range from being clast supported to matrix supported, with clasts ranging from boulder to granule in size. Granitoid and iron formation clasts are common within many of the conglomerate horizons. Wackes locally display partial Bouma cycles, so perhaps these rocks have a turbiditic origin.

The transition from the coarser-grained clastic rocks in the north to the finer-grained clastic rocks in the south is marked by a lineament which may be a fault. Rocks of this assemblage are generally metamorphosed to greenschist grade but are metamorphosed to amphibolite grade where they are in direct contact with granitoid rocks of the Pukaskwa batholith.

Deformation within the sedimentary rocks is seen as small-scale folds of bedding, locally overturned bedding and complex facing-direction reversals. The folds are upright, tight to isoclinal, with shallow plunges to both the east and west. A regional foliation is approximately axial planar to these folds. Dimensional lineations (e.g., conglomerate clasts, quartz grains) plunge 35° to 75° to the north on the foliation.

MINERALIZATION

Mineralization in the Mishibishu greenstone belt consists of:

1. gold-bearing quartz veins hosted within regional zones of deformation (Reid et al., in press): Magnacon gold mine in the Mishibishu deformation zone (MDZ) and

the Eagle River deposit in the Eagle River deformation zone (ERDZ)

2. minor quartz-chalcopyrite ± bornite ± hematite veins associated with fault and/or shear zones (Reid et al., in press)
3. quartz-scheelite veins (e.g., Crane Tungsten deposit; Bennett and Thurston 1977)

Gold mineralization is found within all of the lithotectonic assemblages where associated with regional zones of deformation (Heather 1985, 1986). In contrast to the Schreiber–Hemlo greenstone belt to the north and west, no base metal volcanogenic massive sulphide (VMS) occurrences have been found within the Mishibishu greenstone belt.

Schreiber–Hemlo Greenstone belt

The Schreiber–Hemlo greenstone belt trends easterly from near White River in the east, to Schreiber in the west (*see* Figure 12.2). The trends of lithologic units are locally subparallel with the contacts of the Black–Pic batholith (*see* Figures 12.2 and 12.7). The Port Coldwell alkalic complex intrudes and separates the entire greenstone belt into 2 segments. Two supracrustal assemblages have been proposed for the eastern part, the older is the Hemlo–Black River assemblage to the north, the younger is the Heron Bay assemblage to the south. The 2 supracrustal assemblages represent 2 stratigraphic units of rock that have very similar characteristics and are separated by the Lake Superior–Hemlo fault zone (*see* Figure 12.7). These 2 assemblages are compositionally and geochronologically similar to the Wawa and Catfish assemblages, respectively, of the Wawa area. The supracrustal rocks of the western part of the belt are collectively called the Schreiber assemblage. These rocks remain unsubdivided by age and are currently considered to be temporally equivalent to the Hemlo–Black River assemblage (*see* Figure 12.7). Amphibolite facies metamorphism is ubiquitous.

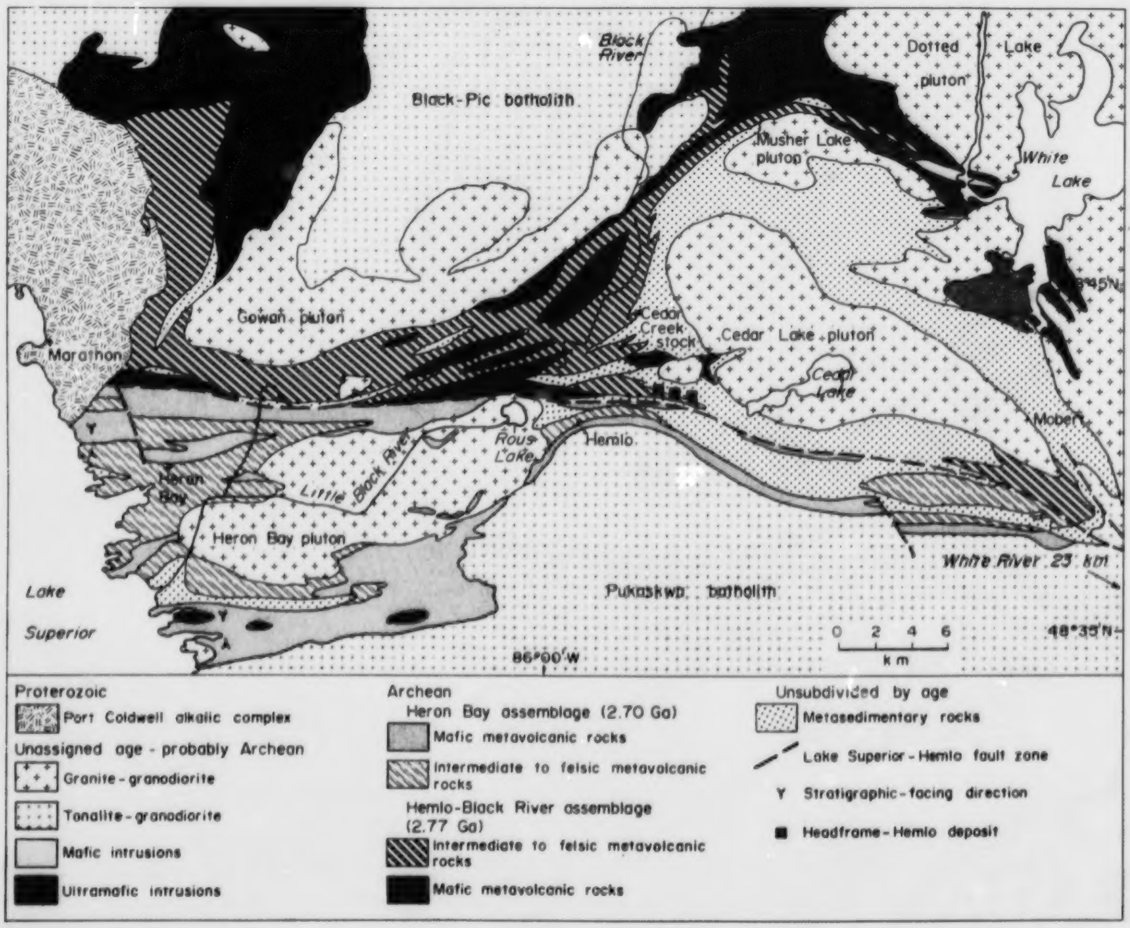

Black River
Dotted Lake pluton
Black-Pic batholith
Musher Lake pluton
White Lake
48°45'N
Gowan pluton
Cedar Creek stock
Cedar Lake pluton
Cedar Lake
Marathon
Mobert
Hemlo
Rous Lake
Little Black River
Heron Bay
Heron Bay pluton
White River 23 km
Lake Superior
Pukaskwa batholith
0 2 4 6
km
48°35'N
86°00'W
Proterozoic
Port Coldwell alkalic complex
Unassigned age - probably Archean
Granite - granodiorite
Tonalite - granodiorite
Mafic intrusions
Ultramafic intrusions
Archean
Heron Bay assemblage (2.70 Ga)
Mafic metavolcanic rocks
Intermediate to felsic metavolcanic rocks
Hemlo-Black River assemblage (2.77 Ga)
Intermediate to felsic metavolcanic rocks
Mafic metavolcanic rocks
Unsubdivided by age
Metasedimentary rocks
Lake Superior - Hemlo fault zone
Stratigraphic - facing direction
Headframe - Hemlo deposit

Figure 12.7. Geological map of the eastern half of the Schreiber–Hemlo greenstone belt (*see* Figure 12.2) showing the structural and lithologic features described in the text.

HEMLO–BLACK RIVER ASSEMBLAGE

A single assemblage is postulated for rocks that extend from White River in the east, along the northern side of the Hemlo greenstone belt, and northwards up into the Black River area (Milne 1968; Muir 1982a; Siragusa 1984a, 1984b, 1985a, 1985b). The southern boundary of the assemblage is the Lake Superior–Hemlo fault zone (*see* Figure 12.7).

Geological mapping has yet to define the precise location of the shear zone-fault system throughout its length, but provisionally, the shear is thought to coincide with chloritic and talcose schists that exhibit high strain along the northern edge of a discontinuous mafic unit. Within the Hemlo–Black River assemblage, layering and foliation strike variably, from east-southeastward in the south, to generally northeastward or northwestward in the north. In the vicinity of Hemlo, structural trends are generally east to east-southeast and the state of strain is high.

Stratigraphic-facing direction determinations are rare and variable; no consistent orientation or structural pattern is clear, especially within sedimentary units and in the highly deformed rocks in the vicinity of Hemlo. In the northwestern part of the assemblage, in the region north and west of Hemlo (*see* Figures 12.2 and 12.7), the supracrustal rocks form a west-facing sequence (Milne 1968). Orientations of layering and foliation are to a large extent governed by the proximity of adjacent granitoid complexes and bodies.

The rocks may be divided into 2 major lithologic groups: diverse tholeiitic and calc-alkalic volcanic rocks, and sedimentary rocks. The volcanic rocks contain significant volcaniclastic equivalents, are perhaps overlain by, or laterally equivalent to a unit of predominantly turbiditic rocks of undetermined provenance (Milne 1968).

Mafic volcanic rocks consisting of pillowed, massive and foliated flows are the lowermost structural unit and contain gabbroic, peridotitic and serpentinous intrusions. Calc-alkalic, mafic to felsic, banded, pyroclastic and volcanic flows succeed the mafic rocks and may be transitional into sedimentary rocks that consist of greywacke, iron formation and conglomerate.

In the western part of the assemblage, volcanic units predominate (Muir 1982b), whereas towards Hemlo, there is an increasing abundance of aluminous sedimentary rocks, particularly north of the Black River (Muir 1982a). Calc-alkalic, pyroclastic and volcaniclastic rocks east of Hemlo *structurally* overlie massive and pillowed tholeiitic volcanic rocks and more highly metamorphosed and deformed equivalents.

Felsic to intermediate volcanic rocks (2772 Ma, Corfu and Muir 1989a) are present in the vicinity of Hemlo, particularly south of the Black River. These volcanic rocks form a calc-alkalic, quartz-feldspar-porphyritic complex (Muir 1986) consisting of pyroclastic deposits and massive flows and/or hypabyssal intrusions. This complex acts as the footwall and, locally, the host to gold mineralization of the large-tonnage gold deposit situated about 5 km east of Hemlo. In the Hemlo area, reworked feldspathic deposits (Muir 1986, 1988, 1989) are genetically associated with the porphyritic complex. At this location, they are the predominant rock type and, especially to the east, are interleaved (interbedded?) with biotite- and hornblende-bearing turbiditic wacke and siltstone units.

Steeply dipping sedimentary rocks within the Hemlo–Black River assemblage virtually surround the Cedar Lake pluton. Layering and tectonic fabrics are subparallel to the granite contact.

The state of strain within the supracrustal rocks ranges considerably. Some units contain well-preserved primary textures and structures; whereas in others, shearing may have all but obliterated the original texture.

Granitoid rocks have intruded the supracrustal rocks and they vary from foliated to gneissic tonalite to granodiorite (e.g., the Black–Pic batholith, Milne 1968). Later, relatively undeformed, discordant granodiorite bodies such as the Cedar Lake pluton (2688 Ma), and the Cedar Creek stock (2684 Ma, Corfu and Muir 1989a) intruded the supracrustal rocks. At the boundary of the Hemlo–Black River assemblage with the Black–Pic batholith, the Gowan pluton (2678 Ma, Corfu and Muir 1989a) forms a crescentic body.

Sulphide mineralization occurs at and near the boundary between the mafic and mafic to felsic volcanic units (Milne 1968). Gold mineralization being mined at Hemlo is located within highly sheared, tectonized, intermediate to felsic volcanic rocks and sedimentary rocks near the southern contact of the assemblage (Burk et al. 1986; Kuhns et al. 1986; Valliant and Bradbrook 1986; Walford et al. 1986).

Previous workers (e.g., Muir 1982a, 1982b; Siragusa 1984a, 1984b, 1985a, 1985b) did not distinguish between the Hemlo–Black River assemblage and Heron Bay assemblage. On the basis of 2 age determinations, Corfu and Muir (1989a) suggested the presence of a major structural discontinuity, possibly represented by the Hemlo fault, between now juxtaposed assemblages. This subdivision is still tentative because of the similar lithologic and structural characteristics of the rocks to the north and south of the Hemlo fault. The limited geochronologic data indicate 2 separate episodes of supracrustal formation in this region. The combination of the regionally extensive zone of shearing at the Lake Superior–Hemlo fault zone, and the geochronologic information, are the authors' justification for the establishment of this assemblage boundary within the greenstone belt.

HERON BAY ASSEMBLAGE

The Heron Bay assemblage (*see* Figures 12.2 and 12.7) has a northern, sheared boundary against the Hemlo–Black River assemblage, separated by the Lake Superior–Hemlo fault zone. The southern boundary is a sheeted, transitional, north-dipping contact with gneisses and granitoid rocks of the Pukaskwa batholith. The Pukaskwa gneissic complex has been dated at 2719 and 2688 Ma (Corfu and Muir 1989a). In the west, the rocks are cut by the composite Port Coldwell alkalic complex, which has been dated at 1108 and 1099 Ma (Heaman and Machado 1987).

The Heron Bay assemblage may be grouped into 2 lithologic divisions: one composed predominantly of tholeiitic, mafic volcanic rocks; the other, predominantly of calc-alkalic, intermediate to felsic, pyroclastic rocks and volcaniclastic equivalents (Milne 1968; Muir 1982b; Siragusa 1984a, 1984b, 1985a, 1985b). The relationship of the tholeiitic rocks with those rocks of calc-alkalic composition is undetermined.

The mafic volcanic rocks are steeply to moderately dipping, iron-rich tholeiites. Plagioclase-porphyritic and/or variolitic flows are locally mappable units. The plagioclase phenocrysts are useful strain markers, being relatively undeformed in the Heron Bay area, and becoming increasingly deformed towards Hemlo. Lenses of ultramafic rocks, including peridotite, hornblendite and their deformed and metamorphosed equivalents, occur sporadically within the tholeiites from Lake Superior to somewhat east of Hemlo (Muir 1982b). Pan and Fleet (1989) showed that some of the ultramafic rocks have a komatiitic composition, but their origin is still obscure.

The calc-alkalic volcanic rocks appear to thicken towards Heron Bay, but it is not clear whether this is an original feature or perhaps one due to strain. In the immediate area of the Lake Superior coastline, a number of east-trending reclined folds have been inferred on the basis of facing reversals. These calc-alkalic volcanic rocks are steeply dipping and strike easterly, bifurcating around the Heron Bay pluton. Diverse, feldspar- and quartz-feldspar-porphyritic, dacitic and rhyolitic flows, and associated minor intrusions are locally interbanded with minor, massive to pillowed, iron-rich tholeiitic flows. Minor quantities of calc-alkalic mafic rocks, including pillow breccia and flows, occur locally within the pyroclastic pile along with interflow mudstones. Some mudstones have climbing ripple structures. Deposits of wacke and siltstone, locally highly folded and transposed, and similar to those in the Hemlo assemblage, occur along the southern contact of the Heron Bay assemblage and within it.

In the Heron Bay area, metamorphic grades are low and strain states are low to moderate, except in sinuous narrow zones that are roughly parallel to the tectonized lithologic layering. In the Hemlo area, especially between the Cedar Lake pluton and the Pukaskwa batholith, strain states and the metamorphic grade are much higher. The metamorphic grade is medium with pressure up to 0.6 GPa (Pan 1990). The folded and transposed volcanic and associated sedimentary rocks (2695 Ma, Corfu and Muir 1989a) were metamorphosed prior to, and during, the emplacement of the Heron Bay pluton (2688 Ma, Corfu and Muir 1989a) and then cut by lamprophyre and diabase dikes.

STRUCTURE OF THE HEMLO AREA

In the Hemlo area (*see* Figure 12.7), at least 2 major stages of folding have been recognized along with 2 subordinate stages (Muir and Elliott 1987). The earliest structural features are rare layer-parallel breccia and associated isoclinal folds, D_1 (Muir and Elliott 1987), that are affected by a regional folding event, D_2. D_1 deformation is inferred to have developed as a result of slides or thrusts that form along surfaces subparallel to the original layering. D_2 deformation produced S-style, asymmetric (F_2) folds, generally pervasive elongation and mineral lineations that are parallel to the fold axes. The orientation of linear structural features (L_2) ranges from shallowly west plunging at Heron Bay, to moderately to steeply northwest plunging at Hemlo, to shallowly east plunging southeast of Hemlo. D_2 deformation appears to have predated, or accompanied, the intrusion of stocks and plutons in the area.

Postdating the D_2 event is a major stage of dextral-sense ductile shearing, which accompanied or postdated the alteration and mineralization related to the Hemlo gold deposit. This ductile shearing produced Z-style (F_3) asymmetric folds (Muir and Elliott 1987).

The Lake Superior–Hemlo fault zone is a zone of intense strain which extends from the Port Coldwell alkalic complex past, and to the south of, the Hemlo gold deposit. The fault zone possibly extends as far east as the town of White River. In places, the shear zone is imperfectly delineated because of poor exposure. Dikes (?) of picritic basalt composition, and of unknown age, occupy the fault zone southeast of Hemlo. The dikes (?) have been converted to lenses of actinolite-chlorite-talc schist. Barite lenses are located along or near the fault, but their origin is obscure (Patterson 1984; Burke et al. 1986).

SCHREIBER ASSEMBLAGE

A belt of supracrustal rocks (*see* Figure 12.2) extends from the Port Coldwell alkalic complex (1108 and 1099 Ma, Heaman and Machado 1987), westwards towards Terrace Bay and Schreiber (Walker 1967; Carter 1988a; Williams 1989) and northwards to Big Duck Lake (Pye 1964).

Three major rock types have been recognized (Carter 1988a): 1) mafic volcanic rocks of tholeiitic composition; 2) calc-alkalic mafic to felsic volcanic rocks; and 3) sedimentary rocks. The first 2 rock types are spatially and perhaps genetically associated, with mafic rocks dominant over felsic. There is, however, no unequivocal stratigraphic succession from mafic to felsic compositions.

Tholeiitic volcanic rocks are pillowed, massive or foliated flows, tuffs and fragmental rocks, and are intruded by pods and lenses of gabbro and peridotite. Felsic fragmental rocks range in clast size from coarse to fine, and structurally underlie the predominantly mafic rocks that are located in the northern part of the belt at Winston Lake and in the vicinity of Schreiber (*see* Figure 12.2). Sedimentary rocks probably overlie the volcanic rocks and are composed of greywacke turbidite, interlayered slate and iron formation (Carter 1988a; Schnieders 1987); they are considered to postdate volcanism (Walker 1967), but evidence is sparse because stratigraphic contacts are commonly sheared.

Stratigraphic-facing determinations are rare; those recorded indicate considerable variation which has been ascribed to folding (Walker 1967). However, within sedimentary units, the complexity of folding is in contrast to the simple, homoclinal structures found within the mafic volcanic rocks. The occurrence of complex facing patterns, lack of regional-scale fold closures and ubiquitous sheared contacts between rock types, render a simple upright folded structure unlikely (Williams 1989). Williams (1989) also suggested that tectonic juxtaposition of units occurred prior to a later stage of upright folding. On a regional scale, east-southeasterly plunging upright folds have been interpreted by Carter (1988a) on the basis of broadly homogeneous zones of opposing facing direction. The Crossman batholith (*see* Figure 12.2) is emplaced into, and perhaps postdates, one such anticlinal fold.

Carter (1988a) considered the calc-alkalic metavolcanic rocks to stratigraphically underlie mafic metavolcanic rocks in the Terrace Bay–Schreiber area. The contact between them is a sulphide-bearing graphitic shale, and by analogy with a similar contact studied in the Jackfish region to the east (Schnieders 1987), it is possibly a tectonic contact. A tectonic contact (Williams 1990) consisting of mylonitized gabbro and felsic rock, exhibits isoclinal folds and strong linear fabrics. This contact separates north-facing gabbroic rocks, associated mafic volcanic rocks, minor felsic volcanic rocks and sedimentary rocks occurring in the Big Duck Lake area (Pye 1964) from a structurally underlying, east-facing felsic volcanic centre at Winston Lake (Severin and Balint 1985). This contact also separates a northern tectonic unit of gabbro and mafic to felsic volcanic rocks from a southern assemblage of predominantly mafic volcanic rocks. Insufficient data on a regional scale preclude elevating these units to assemblage status.

Grades of metamorphism are variable; background levels are of upper greenschist facies, but locally an increase to upper amphibolite facies is recognized around late granitoid plutons (Walker 1967).

Mineralization in the Schreiber assemblage consists mainly of a major zinc deposit at Winston Lake (Severin and Balint 1985), where sphalerite mineralization occurs at the tectonic contact of a layered gabbro and an underlying hydrothermally altered felsic volcanic centre. In the vicinity of Schreiber, sulphidic gossans occur within graphitic shale, which lies between the lower calc-alkalic unit and the structurally overlying, predominantly mafic unit (Carter 1988a). Two abandoned gold mines (Empress and North Shore mines) are located within contact alteration zones associated with hydrous emanations from the Terrace Bay pluton (Marmont 1983). Garnet- and amphibole-rich alteration and minor sulphide mineralization occur within highly sheared rocks along the northern boundary of this assemblage (Williams 1989). This alteration is comparable to the alteration described from Manitouwadge.

Dayohessarah–Kabinakagami Greenstone Belts

The Kabinakagami (Siragusa 1972; Wilson 1989, 1990) and Dayohessarah (Fenwick 1967) greenstone belts are probably 2 separate parts of a once continuous, sinuous supracrustal belt now joined by a trail of amphibolite inclusions within enclosing granitic rocks (*see* Figure 12.2). No internal subdivision of the greenstone belts has yet been possible, so the combined belts are here defined as a combined Dayohessarah–Kabinakagami assemblage. The authors consider this assemblage to be equivalent in age to the Schreiber–Hemlo assemblage on the basis of their close spatial association and apparent continuity.

DAYOHESSARAH–KABINAKAGAMI ASSEMBLAGE

This assemblage consists of poorly preserved, steeply dipping and strongly sheared, upper greenschist to epidote-amphibolite facies, mafic volcanic and clastic sedimentary rocks. This assemblage may correlate with the Manitouwadge–Hornepayne assemblage located to the north (*see* Figure 12.2). Foliations strike in a range from north to east, and the rocks furnish no reliable facing directions (Wilson 1989). The assemblage boundaries are all against granitoid rocks.

The supracrustal rocks are chiefly composed of mafic volcanic rocks and subordinate quartz arenite, wacke, and rarely, conglomerate. Felsic volcanic rocks are rare and are difficult to distinguish from concordant sheets of deformed, intrusive granitoid. Coarse, clastic, felsic volcanic rocks are found in the eastern part of the Kabinakagami belt but are the only unequivocal example.

All supracrustal rocks are marginally sheeted and enclosed by bodies of gneissic to foliated tonalite, trondhjemite, granodiorite and granite that have not been subdivided in detail. Foliated intrusions compositionally consist of tonalite to granodiorite, are intrusive into the supracrustal rocks, and locally produce migmatites. Later intrusions tend to be massive and granodioritic to granitic in composition.

The supracrustal rocks, especially sedimentary rocks, are highly strained, as prominent foliation directions parallel the lithologic layering. Volcanic rocks, rarely containing pillow structures, are elongate parallel to steeply plunging lineated amphibole. Sedimentary rocks are commonly folded on an outcrop scale, but there is no unequivocal evidence that the belt is a synform. A dextral shear zone is located along the central axis of the belt, which also coincides with the change from upper greenschist facies rocks to the northwest to amphibolite facies rocks to the southeast (Wilson 1989).

Manitouwadge–Hornepayne Greenstone Belt

This discontinuous belt of supracrustal rocks forms the northern margin of the Wawa Subprovince. Only 1 assemblage has been recognized (*see* Figure 12.2). The age and correlation of this assemblage with other assemblages is equivocal, but the authors consider it to be equivalent to the Hemlo–Black River assemblage based on their apparent continuity around the northwestern part of the Black–Pic batholith.

MANITOUWADGE–HORNEPAYNE ASSEMBLAGE

A generally east-trending, variably dipping, highly deformed and metamorphosed belt of supracrustal rocks, ranging in width from less than 1 km to more than 2 km, may be traced as a series of discontinuous units from the Lepage fault zone in the east (Berger 1986), through Hornepayne, to west of Manitouwadge (*see* Figure 12.2). The assemblage is enclosed on its south side by the foliated to gneissic, tonalite to granodiorite of the Black–Pic batholith. It is enclosed on the north side by the migmatitic metasedimentary rocks of the Quetico Subprovince (Williams and Breaks 1989). Based on the presence of inclusion trains, the Manitouwadge–Hornepayne assemblage may initially have been continuous with the supracrustal rocks of the northern edge of the Hemlo–Black River assemblage, which is part of the Schreiber–Hemlo greenstone belt. The Manitouwadge–Hornepayne assemblage is lithologically and structurally similar to the Dayohessarah–Kabinakagami assemblage to the south, with which it may originally have been connected.

In the vicinity of Manitouwadge, the supracrustal rocks consist of mafic volcanic, subordinate felsic volcanic and sedimentary rocks, as well as layered gabbro-anorthosite intrusions (Milne 1968; Giguere 1972; Bakker et al. 1985; Williams and Breaks 1989, 1990). Along strike to the east, sedimentary rocks and felsic volcanic rocks are only sporadically present. There is an eastward increase in the proportion of the layered mafic intrusions that contain significant units of gabbroic, ultramafic and anorthositic composition.

Stratigraphic facing has only rarely been determined from mafic to felsic transitions, pillow shapes in mafic volcanic rocks, and graded phase layering of ultramafic to anorthositic compositions within the gabbroic bodies. Stratigraphic facing is generally to the north except within the Manitouwadge synform, where highly deformed mafic and felsic volcanic rocks and sedimentary rocks appear to be repeated by isoclinal folding and then refolded into a northeasterly trending synform (Williams et al. 1990). The belt contains the Manitouwadge copper-zinc volcanogenic massive sulphide deposits (Williams et al. 1990).

The belt is affected by several deformation events, starting with 2 stages of layer-parallel shearing during, or just prior to, the emplacement of the bordering tonalites. These deformation events produced gneisses. Contacts between supracrustal rocks and bordering tonalitic gneisses are commonly mylonitic. All rocks were then deformed into northeast-plunging, upright asymmetrical folds, of which the Manitouwadge synform is the most clearly defined (*see* Figure 12.2).

Chapleau Block

An elongate block of highly metamorphosed supracrustal and intrusive rocks, now gneisses and migmatites, in the Kapuskasing Structural Zone (Thurston et al. 1977; Percival 1981; Percival and Card 1985) forms the north-northeast-trending eastern edge of the Wawa Subprovince (*see* Figure 12.2). The eastern boundary of this block is the Ivanhoe Lake cataclastic zone (Bursnall 1989; Bursnall and Moser 1989). The western margin is poorly defined. Percival and Card (1985) considered the Chapleau block to be transitional into the low-grade granite-greenstone terrane of the Michipicoten greenstone belt. Alternatively, it may be separated from the rest of the Wawa Subprovince by the traces of the Wakusimi and Lepage faults (*see* Figure 12.2; Thurston et al. 1977; Leclair 1990).

The Chapleau block (*see* Table 12.2) consists of granulite and upper amphibolite facies gneisses and foliated rocks, which can be variously assigned sedimentary, volcanic or igneous protoliths. On the Tectonic Assemblages of Ontario map (*see* Map 2576, map case), the authors have interpreted the volcanic rock, and some sedimentary rocks, to be laterally equivalent to the northern group of assemblages assigned an age of about 2.77 to 2.75 Ga. This age is in accord with preliminary ages obtained by Percival and Krogh (1983). The original contact and relative age relationships between the supracrustal and intrusive rocks within the Chapleau block are unknown because of the obliterating effects of deformation and metamorphic recrystallization. Mafic and intermediate granulites, possibly of volcanic derivation, are intruded by the Shawmere Anorthosite Complex, which is a folded sill-like body that contains rocks with a highly calcic plagioclase.

In the northern part of the block, metasedimentary migmatites displaying upper amphibolite to granulite facies mineral assemblages, are transitional northwards into the Quetico Subprovince (*see* Williams and Breaks 1990; *see also* Williams, this volume). All these rock types, in turn, have been intruded to varying degrees, during and after deformation, by tonalitic to granitic sheets and plutons.

Foliation trends within the Chapleau block describe a shallow arc, from north-northeast trends in the east, to nearly easterly trends in the south and west (Moser 1989). Dip values are variable, usually moderate to the northwest. The pattern of foliation orientations and the location of lithologic units may be interpreted as being due to regional distortion of easterly trending units by sinistral ductile shear along the eastern margin of the Kapuskasing Structural Zone. However, areas close to this eastern boundary, adjacent to the Ivanhoe Lake cataclastic zone, exhibit steep westerly dips and equivocal kinematic indicators (Bursnall and Moser 1989).

Regions of Granitoid Rock Between Major Greenstone Belts

Relatively little is known of the vast areas of granitoid rock that separate the major groups of greenstone belts. These large masses are known to be compositionally and structurally inhomogeneous, based on a review of the field mapping, aeromagnetic and aeroradiometric data. Broad subdivision (Card 1979, 1982) of the granitoid masses into older tonalitic to granodioritic and younger masses of granodioritic to granitic compositions has been attempted. These subdivisions are shown relatively unmodified on the east-central sheet, Bedrock Geology of Ontario map (*see* Map 2543, map case).

In accord with data from the Wabigoon Subprovince, the authors consider that the tonalitic rocks may, in part, be synvolcanic with regard to the 2.77 and 2.70 billion-year-old greenstone assemblages. Therefore, on the Tectonic

Assemblages of Ontario map (*see* Map 2577, map case), the tonalites are noted as having an age range from as old as the Wawa and Hemlo assemblages (2.77 Ga) to as young as the youngest granites (2.67 Ga), but the preferred age is about 2.7 Ga, with the granitic rocks being slightly younger.

Both foliated to massive, and massive to gneissic rocks occur in regions between major greenstone belts and included in them are many relicts, enclaves or septa of supracrustal rock, only some of which may be traced directly into substantial greenstone belts. More detailed work (e.g., Percival 1981; Moser 1988) has shown these subdivisions to be more complex at a small scale, but the rocks are insufficiently exposed, or insufficiently geophysically distinct to enable precise mapping of boundaries between igneous phases. The following sections describe what detail is available and serve to demonstrate how little is known of the most abundant rock in the Superior Province.

PUKASKWA BATHOLITH

The Pukaskwa batholith (cf., Pukaskwa Gneissic complex, Corfu and Muir 1989a) forms the southern border of the Hemlo greenstone belt, and the northern border of the Mishibishu and Michipicoten greenstone belts. At Hemlo, rocks from the margin have been dated at 2719 and 2688 Ma. At the southern flank of the batholith, north of the Mishibishu greenstone belt, tonalite has been dated at 2721 Ma. Bennett and Thurston (1977) described the batholith as consisting, in part, of migmatitic and gneissic zones in which supracrustal remnants are common. They also described the composition of the batholith as consisting, in part, of leucotonalite (trondhjemite) to quartz monzonite. Patches and sheets of pegmatite and aplite are ubiquitous.

BLACK–PIC BATHOLITH

The Black–Pic batholith was described by Milne (1968) as a foliated to massive rock of tonalitic to granodioritic composition. The batholith is a structural dome with foliation dips that are shallow to moderate outward from the centre (Williams and Breaks 1989, 1990). Massive variants of gabbro, diorite, tonalite, granite and granodiorite are considered to be late-stage intrusions through the domal structure. On the northern margin of the Black–Pic batholith, the Manitouwadge–Hornepayne greenstone belt is structurally concordant with tonalitic gneisses marginal to the batholith, which is, therefore, considered to have been emplaced some time after 2.77 Ga. At the margin of the Black–Pic batholith with the Manitouwadge–Hornepayne greenstone belt, and elsewhere, mafic metavolcanic rocks have hybridized with the tonalite to form melanocratic tonalite and diorite in which inclusions of mafic origin are numerous.

In the central part of the Black- Pic batholith, to the south of the Kabinakagami–Dayohessarah greenstone belt, Tortosa (*in* Bennett et al. 1987) delineated a number of igneous phases, including a tonalitic dome surrounded by supracrustal septa which, in detail, is reminiscent of the much larger-scale domal structure seen between the Hemlo and Manitouwage greenstone belts.

WAWA GNEISS TERRANE

Mapping (Moser 1988; Leclair and Nagerl 1988; Leclair 1990) in rocks previously considered intractable shows the Wawa gneiss terrane to be characterized by massive to foliated tonalite and granodiorite (*see* Figure 12.2). Occurring locally, are domal structures in tonalitic gneisses, tracts of mafic gneiss of unknown protolith, high-strain zones and late-stage granitic rocks. A number of intrusive events involving tonalite, diorite, gabbro, granodiorite and granite are recognized by Moser (1988). Shallowly east-plunging mineral and stretching lineations occur on generally west-northwest-striking gneissic layering. The mineral and stretching lineations are considered to represent the effects of extensional tectonics within these high-grade rocks (Moser 1988). The Wawa gneisses are considered by some (e.g., Percival 1990) to be representative of mid-crustal intrusions that have developed during, or after, the production of greenstone belts in the late Neoarchean.

Saganagons and Shebandowan Greenstone Belts

INTRODUCTION

Greenstone belts in the westernmost part of the Wawa Subprovince in Ontario (*see* Figure 12.2) extend from Lake Superior southwestward to Minnesota. In this section, the geology of 2 greenstone belts that border a central batholithic complex is described. The Shebandowan greenstone belt forms the northern margin of this subprovince, separating the Quetico Subprovince in the north from the Northern Light–Perching Gull Lakes (NLPG) batholitic complex in the south (*see* Figure 12.2). The Saganagons greenstone belt lies *en échelon* from the Shebandowan greenstone belt, abuts the NLPG batholitic complex and continues into the Vermilion district of northeastern Minnesota. Archean rocks located to the southeast and east of the subprovince, near Lake Superior, are overlain with marked angular unconformity by flat-lying Proterozoic sedimentary rocks of the Animikie Basin, and diabase sills of the Nipigon Embayment.

GENERAL GEOLOGY AND GEOLOGICAL SUBDIVISION

Most of the Shebandowan area has received detailed geological mapping. Although comments were made by early geologists such as Coleman (1896) and McInnes (1899), the first systematic mapping began with Tanton's (1938) map. The most recent maps in the western half of the belt were produced by Giblin (1964), Hodgkinson (1968), Harris (1968, 1970), Morin (1973) and Srivastava and Fenwick (1973). The eastern half of the belt received attention in the 1980s from Carter (1988b) and references therein, Carter (1990a, 1990b, 1990c) and Scott (1990).

The 2 greenstone belts of the Shebandowan area, the Shebandowan and the Saganagons, may once have been continuous. Northeast of the Saganagons belt, amphibolitic inclusions in granitic plutons and gneisses probably comprise remnants of a previously more continuous supracrustal domain that included the Saganagons and Shebandowan greenstone belts. These greenstone belts are subdivided into assemblages as shown in Figure 12.8. Granitoid rocks from this region are described after having dealt with the structural development. The deformation pattern across the subprovince indicates that the greenstone belts were subjected to early deformation during belt assembly, followed by late penetrative deformation that was confined, generally, close to the Quetico Subprovince. These ductile deformation events were followed by the production of a conjugate set of faults.

SHEBANDOWAN GREENSTONE BELT

The Shebandowan greenstone belt is arcuate in shape (*see* Figure 12.2), reflecting the pronounced curvature in the Quetico–Shebandowan greenstone belt boundary. This shape complements the narrowing of the Quetico Subprovince directly north of the belt. The necking or pinching of the Quetico Subprovince is the most dramatic example of several such arcuate subprovince boundaries observed in the Superior Province; for example, the narrowing of the English River and Winnipeg River subprovinces, north of Savant Lake (*see* Breaks , this volume; *see also* Beakhouse, this volume).

Subdivision of the Shebandowan greenstone belt into 2 oppositely facing assemblages and a third, younger, unconformably overlying assemblage that straddles the first 2, is based on a regional correlation of marker units, contrasts in rock types between assemblages and a compilation of local top or way-up directions. The boundary between the older Greenwater and Burchell assemblages is only approximately defined and may be a fault (*see* Figure 12.8). Further subdivision of these older assemblages into volcanic cycles is also provisional.

The volcanic cycles each typically consist of a lower sequence of tholeiitic basalt flows and an uppermost sequence of calc-alkalic andesite, dacite and rhyolite. For some cycles adjacent to the belt margins, only the lower basaltic part is preserved. Particularly notable in this volcanic belt are the abundance of mafic sills and a prominent, differentiated gabbroic-anorthositic pluton. The stratigraphic units in this belt are near-vertically dipping. The following descriptions focus on the key distinguishing features; the main characteristics of each assemblage are summarized in Table 12.3.

Two contrasting suites of volcanic and sedimentary rocks occur in this belt: 1) an older suite of mafic to felsic volcanic cycles that consist of tholeiitic to calc-alkalic rocks and some komatiitic units; and 2) a younger, unconformably overlying suite of sedimentary and volcanic rocks, including units of alkalic affinity, that resemble rocks of the Timiskaming group (Cooke and Moorhouse 1969) near Kirkland Lake in the Abitibi belt.

The older suite comprises 2 assemblages that stratigraphically face away from each other. They are 1) the Burchell assemblage in the north, consisting of 3 northward younging volcanic cycles; and 2) the Greenwater assemblage in the south, consisting of 3 southward younging cycles.

The younger suite of rocks, defined as the Shebandowan assemblage, consists of sedimentary and volcanic rocks that unconformably overlie, and locally straddle, the boundary between the Burchell and Greenwater assemblages. The main characteristics of these 3 assemblages are summarized in Table 12.3. Regional correlations of assemblages proposed in this chapter have been extended to include the supracrustal stratigraphy in the Vermilion district of Minnesota.

SAGANAGONS BELT

The Saganagons belt is similarly composed of volcanic cycles of tholeiitic basalt flows and units of calc-alkalic dacite and rhyolite; but these cycles are separated into 2 assemblages, the Saganagons and Greenwater assemblages, which face each other across the Knife Lake fault system (*see* Figure 12.8). The Saganagons assemblage is in fault contact on its south margin with epiclastic and pyroclastic rocks of the younger Knife Lake assemblage (*see* Table 12.3 and Figure 12.8).

The Northern Light–Perching Gull Lakes (NLPG) batholithic complex comprises several granitoid plutons, gneisses and megainclusions of volcanic rocks, principally amphibolite (Percival 1983).

The following descriptions of these belts use the assemblage subdivisions as the basis for stratigraphic synthesis. Only the Shebandowan greenstone belt has received significant mineral exploration. It is also the focus for the interpreted sequence of tectonic events in this region.

The juxtaposition of the Saganagons and Greenwater assemblages that face each other across the Knife Lake fault system has been complicated by left-handed strike-slip displacement along this system. On the basis of the geometry of fault branching, the fault system appears to have widened a part of the belt by strike-slip duplexing, notably southwest of Saganagons Lake. Regional stratigraphic relations allow correlation of the greenstone wedge, located northwest of the Knife Lake fault system, with the Greenwater assemblage in the Shebandowan greenstone belt (*see* Greenwater Assemblage). The greenstone wedge northwest of the Knife Lake fault system may also correlate with the Newton Lake Formation of the Vermilion district in northeastern Minnesota (Sims 1976; Green and Schulz 1982). The southeastern block, named the Saganagons assemblage, projects eastward towards supracrustal remnants in the NLPG batholitic complex. To the southwest, the Saganagons assemblage pinches out and is transected by a fault, bounding the northern margin of the younger Knife Lake assemblage of epiclastic and volcaniclastic units. Based on the makeup of dominantly tholeiitic basalts and long stratigraphic markers of iron formation, the Saganagons assemblage compares with the northward-younging upper member of the Ely greenstone belt in Minnesota (Schulz 1980).

Individual tectonic assemblages that together make up the 2 greenstone belts are now described. The principal features of the assemblages are summarized in Table 12.3.

BURCHELL ASSEMBLAGE

Cycle 1 of the Burchell assemblage (*see* Figure 12.8) is intruded by the Shebandowan Lake pluton and continues eastward to Lake Superior, where it is partly covered by the Timiskaming assemblage. The lower part of this cycle comprises massive basalt flows that are locally altered to chlorite schist. Pillow lava underlies the dacite and rhyolite units in the upper part of the cycle, located north of Lower Shebandowan Lake. Near the base of this cycle are tabular, serpentinized, peridotite units that contain copper-nickel mineralization. Such units host the ore deposit of the Shebandowan Mine (Lavigne, Aubut and Scott 1990). The upper part of this cycle is marked by discontinuous units of massive and pyroclastic dacite and rhyolite, possibly reflecting separate volcanic centres (*see* Figure 12.8). The coarsest rhyolitic, lithic breccia occurs west of the Myrt Lake batholith.

Cycle 1 contains the most extensive basic intrusions in the greenstone belt, including a large mafic pluton and thick sills that extend southwestward from Upper Shebandowan Lake. Hodgkinson (1968) has shown that the dominantly gabbroic pluton, and other sill-like intrusions in the north half of this belt, comprise hornblendite-hornblende gabbro-anorthosite. Compositional banding occurs locally. In common with some volcanic assemblages in other subprovinces (e.g., the Confederation assemblage of the Uchi Subprovince), these assemblages show considerable continuity along strike. It is therefore noteworthy that these mafic intrusions are on strike with the voluminous, layered, mafic intrusions in the upper Newton Lake Formation of northeastern Minnesota (Green and Schulz 1982).

Cycle 2 also contains numerous mafic intrusions, but the most distinguishing feature is a thick unit of rhyolite to dacite that extends along the northern part of the belt. The basaltic base is similar to that of cycle 1, with massive basalt in the west and more complex interlayered units of pillowed and massive basalt and andesite to the east. The upper felsic to intermediate part of this cycle partly reflects the compositional pattern of cycle 1, with more proximal massive flows, vent breccias and ash flows of rhyolitic composition in the west (Giblin 1964; Harris 1970; Thurston 1985), changing to more distal dacitic to andesitic tuff in the east, north of Lower Shebandowan Lake (Morin 1973). This unit, located northwest of the Hood Lake syenite pluton, displays proximal flow breccia flanked by massive, unbedded, rhyolitic ash flows (Thurston 1985). The main mass of this felsic unit, located in and south of the Burchell area, is displaced and separated from the more tuffaceous eastern part by right-handed transcurrent movement on the Crayfish Creek fault (*see* Figure 12.2).

Cycle 3 is represented only by a pillowed to massive basaltic unit with some interbedded magnetite-quartz iron formation. Cycle 3 is located north and west of Burchell Lake.

SAGANAGONS ASSEMBLAGE

In this volcanic arc-like assemblage (*see* Figure 12.8), massive flows, volcanic breccia and tuff units of rhyolite to dacite composition occur intercalated with basalt. The most massive rhyolite and coarsest, angular breccia of probable vent facies are found on the south side of Saganagons Lake where there is also an abundance of feldspar-quartz porphyry dikes.

Pillow shapes (Harris 1968) reveal a consistent northward younging direction for the entire breadth of the section from Saganaga Lake to the Knife Lake fault system. The schistosity generally dips steeply southward toward the Saganaga pluton.

GREENWATER ASSEMBLAGE

The Greenwater assemblage in the Saganagons belt is characterized by pillowed basalt and overlying felsic tuff to lapilli tuff and tuff breccia (*see* Figure 12.8). Part of the boundary between the mafic and felsic volcanic rocks is marked by the presence of an iron formation unit. Thurston (1984) recognized the felsic volcanic unit, which is up to 1 km thick, as being dominantly composed of ash flow overlain by air-fall tuff. Graded bedding in this unit indicates that the assemblage youngs southeastward toward the Saganagons assemblage. The Knife Lake fault system separates the 2 assemblages; part of the fault system comprises steep, imbricate faults that repeat a part of the Greenwater assemblage by stacking the vertical strata to form left-handed strike-slip duplexes in a style described by Woodcock and Fischer (1986).

Cycle 1 of the Greenwater assemblage is a relatively thin unit (400 m to approximately 5 km thick) that can be traced much of the length of the greenstone belt. The base of this mainly southward-younging cycle is composed of massive basalt with few pillowed units. This cycle is juxtaposed against the northward-younging cycle 1 of the Burchell assemblage along a faulted (?) contact. The top of the cycle is dacitic to rhyolitic, distal tuff to tuff breccia. It is less than 500 m thick and can be traced nearly continuously from west of Greenwater Lake to Laurie Township, south of the Highway 11–Highway 17 intersection. The coarsest, angular breccia is limited to a few exposures among ash-flow deposits on the north shore of Greenwater Lake. The breccia is only found within a 700 m thick section. Eastward, this unit comprises felsic tuff breccia and tuffaceous ash flows that are several tens of metres thick (Thurston 1985). This unit gradually fines farther east, and to the west, of Greenwater Lake.

Cycle 2 forms the thickest section of volcanic strata east of Greenwater Lake. The broadest exposed section of the base of this cycle lies east of the Greenwater Lake granodiorite pluton. The base of cycle 2 is characterized by massive basalt flows interbedded locally with peridotitic komatiite flows plus magnetite-quartz and magnetite-jasper iron formation units. Distinctive jasper-bearing iron formation from this assemblage occurs as clasts in conglomerates within the unconformably overlying Timiskaming assemblage. Peridotitic flows and sills, typically tens of metres thick, occur mainly east of Greenwater Lake. The flows display polygonal suturing and spinifex texture (Carter 1984, 1985b). Numerous mafic sills, as well as lesser ultramafic sills, occur in this cycle, as they do elsewhere in the belt.

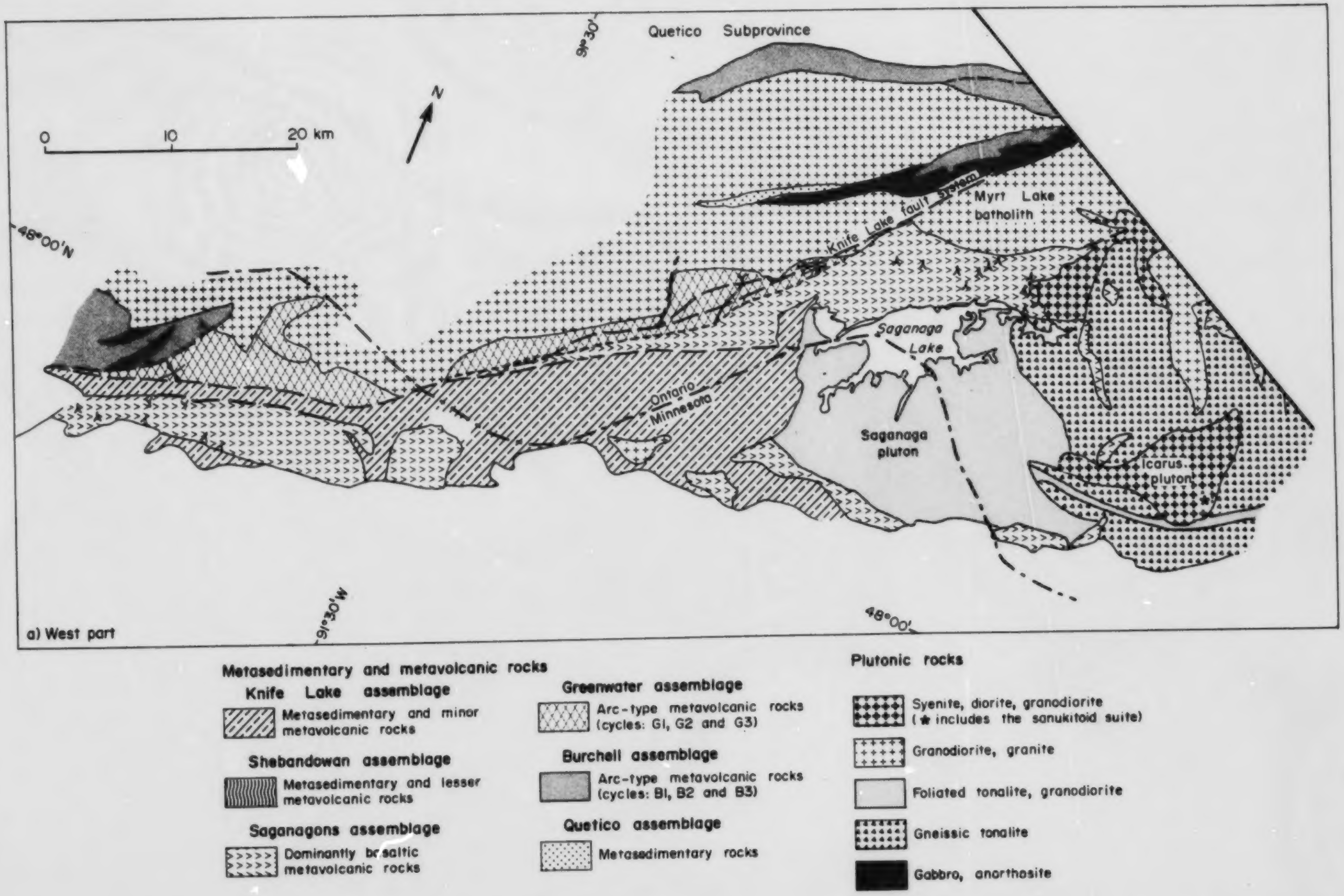

Quetico Subprovince
9°30'
0
10
20 km
N
48°00'N
Knife Lake fault system
Myrt Lake batholith
Saganaga Lake
Ontario
Minnesota
Saganaga pluton
Icarus pluton
91°30'W
48°00'
a) West part
Metasedimentary and metavolcanic rocks
Knife Lake assemblage
Metasedimentary and minor metavolcanic rocks
Shebandowan assemblage
Metasedimentary and lesser metavolcanic rocks
Saganagons assemblage
Dominantly basaltic metavolcanic rocks
Greenwater assemblage
Arc-type metavolcanic rocks (cycles: G1, G2 and G3)
Burchell assemblage
Arc-type metavolcanic rocks (cycles: B1, B2 and B3)
Quetico assemblage
Metasedimentary rocks
Plutonic rocks
Syenite, diorite, granodiorite (★ includes the sanukitoid suite)
Granodiorite, granite
Foliated tonalite, granodiorite
Gneissic tonalite
Gabbro, anorthosite

Figure 12.8a. Supracrustal assemblages and younging directions in the Saganagons (a) and Shebandowan (b) greenstone belts. Younging directions are taken from Ontario Geological Survey maps and reports cited in the text and from Stott (1986) and Kehlenbeck (1988). Also shown are active and abandoned mines. U-Pb zircon age determinations from Corfu and Stott (1986) are numbered as follows: 1) porphyry sill, 2733± 3 Ma; 2) Shebandowan Lake pluton, 2696± 2 Ma; 3) latite volcanic breccia of the Shebandowan assemblage, 2689^{+3}_{-2} Ma; and 4) posttectonic Burchell pluton 2684^{+6}_{-3} Ma.

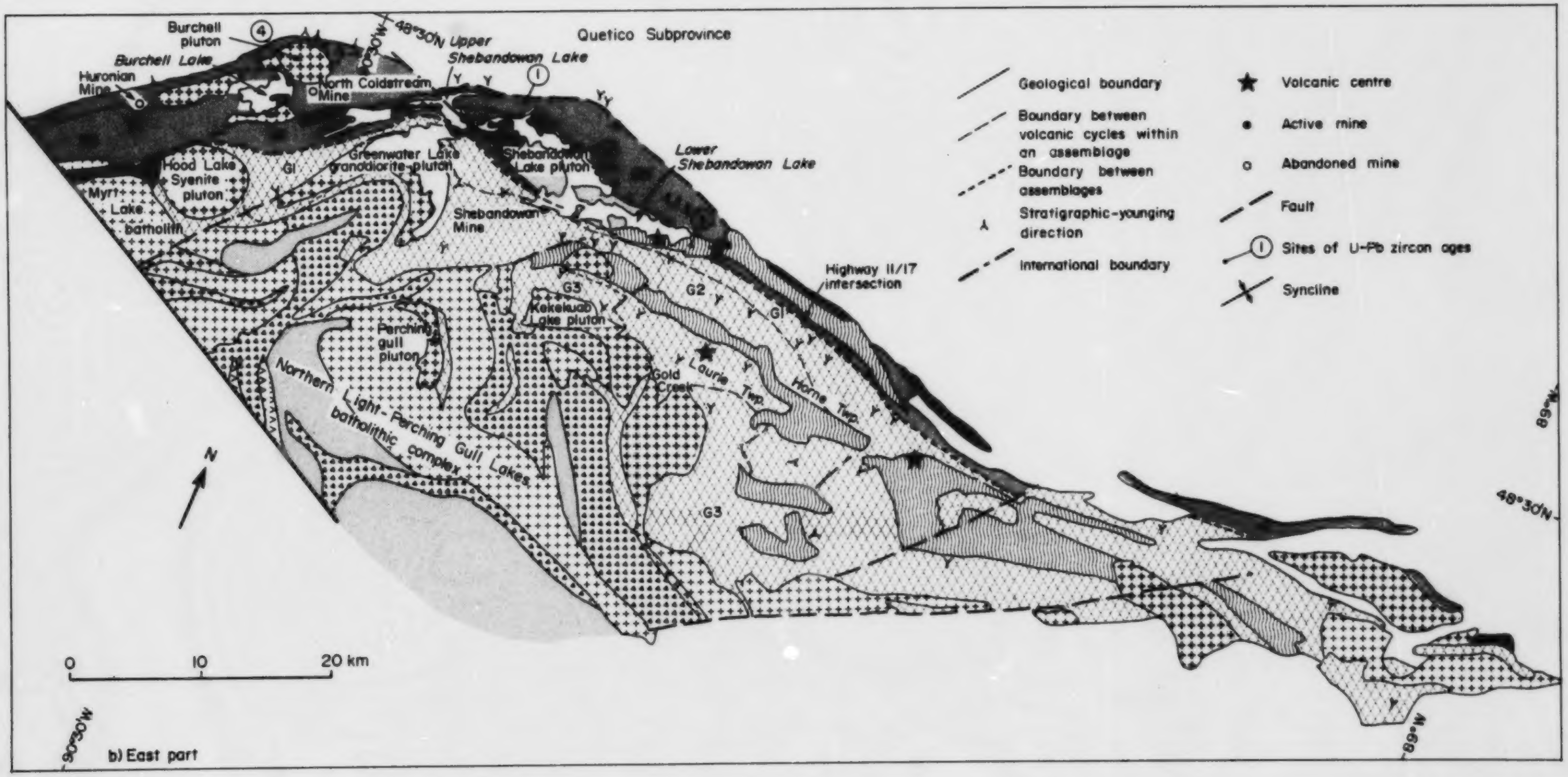

Burchell pluton
④
48°30'N
Upper Shebandowan Lake
Quetico Subprovince
Burchell Lake
Huronian Mine
North Coldstream Mine
①
Hood Lake Syenite pluton
G1
Greenwater Lake granodiorite pluton
Shebandowan Lake pluton
Lower Shebandowan Lake
Myrt Lake batholith
Shebandowan Mine
Highway 11/17 intersection
G3
Kekekuab Lake pluton
G2
G1
Perching gull pluton
Gold Creek
Laurie Twp.
Horne Twp.
Northern Light - Perching Gull Lakes batholithic complex
N
G3
89°W
48°30'N
89°W
90°30'W
0
10
20 km
b) East part
Geological boundary
Boundary between volcanic cycles within an assemblage
Boundary between assemblages
Stratigraphic-younging direction
International boundary
Volcanic centre
Active mine
Abandoned mine
Fault
Sites of U-Pb zircon ages
Syncline

Figure 12.8b. Supracrustal assemblages and younging directions in the Saganagons (a) and Shebandowan (b) greenstone belts. Younging directions are taken from Ontario Geological Survey maps and reports cited in the text and from Stott (1986) and Kehlenbeck (1988). Also shown are active and abandoned mines. U-Pb zircon age determinations from Corfu and Stott (1986) are numbered as follows: 1) porphyry sill, 2733± 3 Ma; 2) Shebandowan Lake pluton, 2696± 2 Ma; 3) latite volcanic breccia of the Shebandowan assemblage, 2689^{+3}_{-2} Ma; and 4) posttectonic Burchell pluton 2684^{+6}_{-3} Ma.

They occur principally across the breadth of the lower, mafic volcanic part of the cycle.

The rhyolitic to dacitic upper part of cycle 2 has been examined locally by Chorlton (1987), who noted that this part of the cycle includes proximal, massive and flow-banded rhyolite, coarse pyroclastic rocks and quartz- and feldspar-phyric intrusive and extrusive rocks. Chorlton also noted the presence of widespread fracturing and brecciation, which encouraged enhanced hydrothermal alteration of these rocks (Chorlton 1987).

Cycle 3 comprises the mafic volcanic suite that lies at the southern margin of the belt. Carter (1990a) described these rocks as massive, fine-grained basalt flows of greenschist to amphibolite grade. They are lithologically similar to the basal part of cycle 2.

SHEBANDOWAN ASSEMBLAGE

The post-Keewatin greenstone, late Neoarchean, Shebandowan intracratonic basin assemblage in the Shebandowan greenstone belt has been recognized for many years as a distinct suite of volcanic and clastic sedimentary rocks (Pye and Fenwick 1965; Shegelski 1980). The Shebandowan assemblage is similar to the Timiskaming assemblage in the Abitibi Subprovince (Cooke and Moorhouse 1969). Tanton (1938) suggested, on the basis of regional map patterns, that an angular unconformity exists between the Shebandowan assemblage and the older, Burchell and Greenwater assemblages (*see* Figure 12.8). No exposures of the unconformity have been observed. However, Borradaile and Brown (1987) have shown that a contrast in structures between these assemblages supports the existence of an unconformity.

Detailed mapping by Carter (1985a) has shown that locally, the strike of the Shebandowan assemblage varies from the underlying older assemblages. Examples are: 1) an unconformity evident from the map patterns in the Horne Township area (*see* Figure 12.8); and 2) along the south side of Lower Shebandowan Lake (Figure 12.9) where a fault contact is reported between the Shebandowan and Greenwater assemblages and where fold axes and stratification of the Shebandowan assemblage units strike at a high angle to the strike of the adjacent Greenwater assemblage (Borradaile and Brown 1987; Carter 1988b). Figure 12.10 also shows unconformable relations on a regional scale in an area farther east. Kehlenbeck (1988) described an outlier of the Shebandowan assemblage north of Thunder Bay that is fault bounded against the adjacent, older Greenwater assemblage. Structural characteristics assist in discriminating these rocks from the older Burchell and Greenwater assemblages and are described *under* Structural Geology.

The Shebandowan assemblage (Shebandowan group of Borradaile and Brown 1987) is a distinctive suite of rocks noted by Shegelski (1980) to consist of: 1) a potassium-rich shoshonitic volcano-plutonic suite of rocks that is in association with calc-alkalic rocks and includes an abundance of massive, unsorted volcanic breccia (Figure 12.11) with reddish, mottled hematite pigmentation; and 2) polymictic conglomerate, arkose, mudstone, and various iron formation units including magnetite-jasper, and graphite-hematite-black chert.

The Shebandowan assemblage is most readily recognized by the association of massive, unsorted debris flows, interbedded with volumetrically dominant sedimentary rocks that include graded tuff, pink and grey arkosic sandstone, and pink and green polymictic conglomerate. Volcanic clasts within these rocks are commonly amphibole-phyric. Sandstone units typically display cross-beds and ripple cross-lamination. The suite of sedimentary rocks is interpreted to represent a fluvial-alluvial fan environment (Shegelski 1980; Carter 1984).

The volcanic suite (Shegelski 1980; Carter 1985b, 1989) contains distinctive red and grey shoshonite and mottled red and green potassium-rich andesite to latite flows (*nomenclature of* Mackenzie and Chappell 1972). The prominence of pyroclastic debris flows and flow breccia that bear subangular, monolithic fragments is evidence of a proximal facies association. Shoshonitic flows, reported by Carter (1984) to be up to 50 m thick, do not show pillow structures. The intimate association of these alkalic volcanic flows with sediments of alluvial-fluvial facies and the pervasive hematitic pigmentation indicate a subaerial environment of deposition for much of this assemblage.

Syenitic stocks and dikes, and mafic dikes (including lamprophyres), are spatially associated with this assemblage east and north of Greenwater Lake (Hodgkinson 1968), and west of the Kaministiquia River (Carter 1985b).

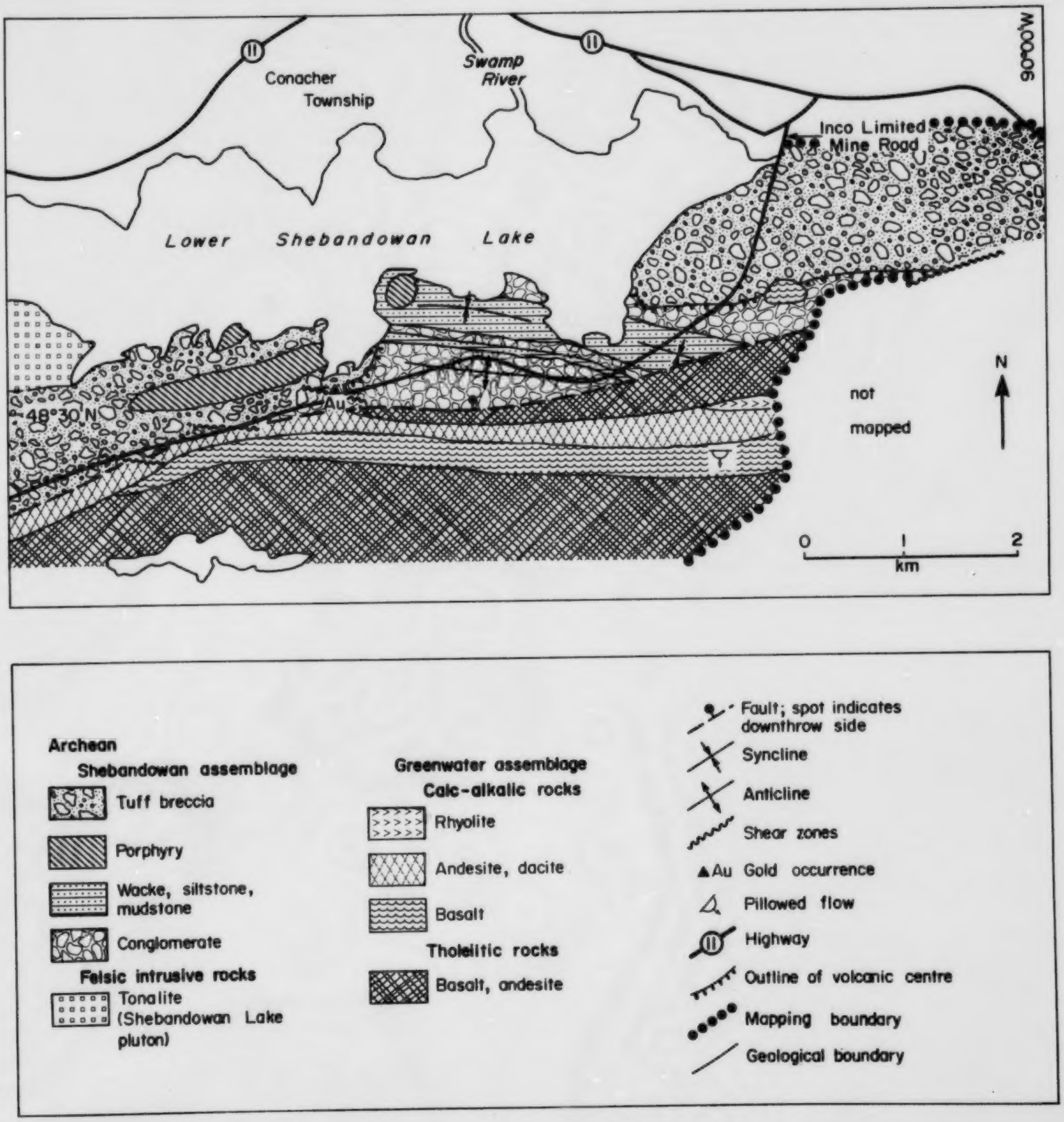
The Shebandowan assemblage is geologically similar to linear, fault-bounded basins that form the depocentres of similar rock groups; for example, in the east Pilbara block of western Australia (Krapez 1984), as well as in the Abitibi greenstone belt of Ontario (Cooke and Moorhouse 1969) and Quebec (Mueller et al., in press). Shegelski (1980) noted that volcanic rocks of the Shebandowan group bear a close similarity to rocks of shoshonite association formed in modern island-arcs during late-stage thickening of the crust. The cratonization of the Shebandowan greenstone belt is conceived to be part of a larger-scale process along the length of each of the greenstone-granite subprovinces. This late timing of volcanism and sedimentation within the Shebandowan assemblage is confirmed by ages obtained by Corfu and Stott (1986).

KNIFE LAKE ASSEMBLAGE

This late sedimentary assemblage (*see* Figure 12.8) comprises the Knife Lake Group of volcanogenic, coarse- to fine-grained clastic sedimentary rocks (Morey et al. 1970; Sims 1972). Only 2 relatively thin wedges of the Knife Lake Group are exposed within Ontario and they are separated by a fault from the main mass of the Knife Lake Group within Minnesota. Nevertheless, these 2 occurrences, shown on Figure 12.8, typify the rock types within the epiclastic part of the Knife Lake Group (Green 1970; Goldich et al. 1972; Ojakangas 1972a, 1979; McLimans 1972). The rocks of this group in Ontario comprise 2 suites: one dominated by wacke turbidite and another dominated by interbedded arkose and conglomerate.

The main suite, as described in Minnesota by Ojakangas (1972a, 1979), consists mainly of volcanogenic wacke, slate and siltstone. There is a mix of sand-sized volcanic lithic fragments, from rhyolite to basalt in composition, in the wacke along with volcanic plagioclase and quartz crystals. The wacke beds are typically graded, interbedded with slate, and also contain sedimentary structures such as convolutions, flame structures and mud-chip zones. These features are typical of resedimented units. In some areas, the turbidites may be resedimented pyroclastic material. There are separate andesitic to dacitic pyroclastic suites in the Minnesota part of the Knife Lake Group, but tuff beds also occur within the sediments.

The second suite is exposed at Cache Bay on Saganaga Lake and overlies the Saganagons assemblage and a lobe of the Saganaga pluton. These sedimentary rocks consist of conglomerate and arkose, comprising one of several occurrences across the Ontario–Minnesota border that cluster close to the Saganaga pluton (Gruner 1941). The



11
Conacher Township
Swamp River
11
90°00'W
Inco Limited Mine Road
Lower Shebandowan Lake
48°30'N
Au
not mapped
N
0
1
2
km
Archean
Shebandowan assemblage
Tuff breccia
Porphyry
Wacke, siltstone, mudstone
Conglomerate
Felsic intrusive rocks
Tonalite (Shebandowan Lake pluton)
Greenwater assemblage
Calc-alkalic rocks
Rhyolite
Andesite, dacite
Basalt
Tholeiitic rocks
Basalt, andesite
Fault; spot indicates downthrow side
Syncline
Anticline
Shear zones
Au Gold occurrence
Pillowed flow
11 Highway
Outline of volcanic centre
Mapping boundary
Geological boundary

Figure 12.9. Fault-bounded contact between folded rocks of the Shebandowan assemblage and rocks of cycle 1 in the Greenwater assemblage of the Shebandowan greenstone belt (*modified from* Morin 1973).

southeastern part of the Cache Bay occurrence consists almost exclusively of detritus eroded from the pluton. The conglomerates contain boulders and cobbles derived from the distinctive quartz-phyric, tonalitic Saganaga pluton.

The matrix of these sedimentary rocks is composed of finer, poorly sorted, texturally immature granitic detritus (McLimans 1972). Some of the boulders at Cache Bay are greater than 3 m in diameter (Goldich et al. 1972). Conglomerate lying on the volcanic belt contains only volcanic detritus, indicating that this sedimentary suite is locally derived. The conglomerate beds are locally overlain by arkose and interbedded with wacke-slate units. Changes in sedimentary structures in the conglomerate occurrences, southwestward away from the Saganaga pluton, are interpreted by Ojakangas (1979) to compare favourably with resedimented conglomerates of a submarine fan. Although part of the conglomerates lie directly on the Saganaga pluton, this suite generally lies within the upper part of the Knife Lake Group. This implies that the unroofing of the Saganaga pluton occurred during the later stages of sedimentation.

MAFIC INTRUSIVE ROCKS

There is an abundance of thick (100 to 600 m), gabbroic, and locally anorthositic and ultramafic, sills in the Shebandowan greenstone belt, most notably intruding the Burchell assemblage. Such prominent gabbroic and related intrusions are not uncommon in belts close to the Quetico Subprovince, such as at Manitouwadge and Winston Lake. Within the Shebandowan greenstone belt is a differentiated hornblendite-gabbro-anorthosite pluton located between Upper and Middle Shebandowan lakes. Anorthositic sills lie within the northern half of the belt amongst thick and coarse-grained intrusions. In the southern half of the belt,

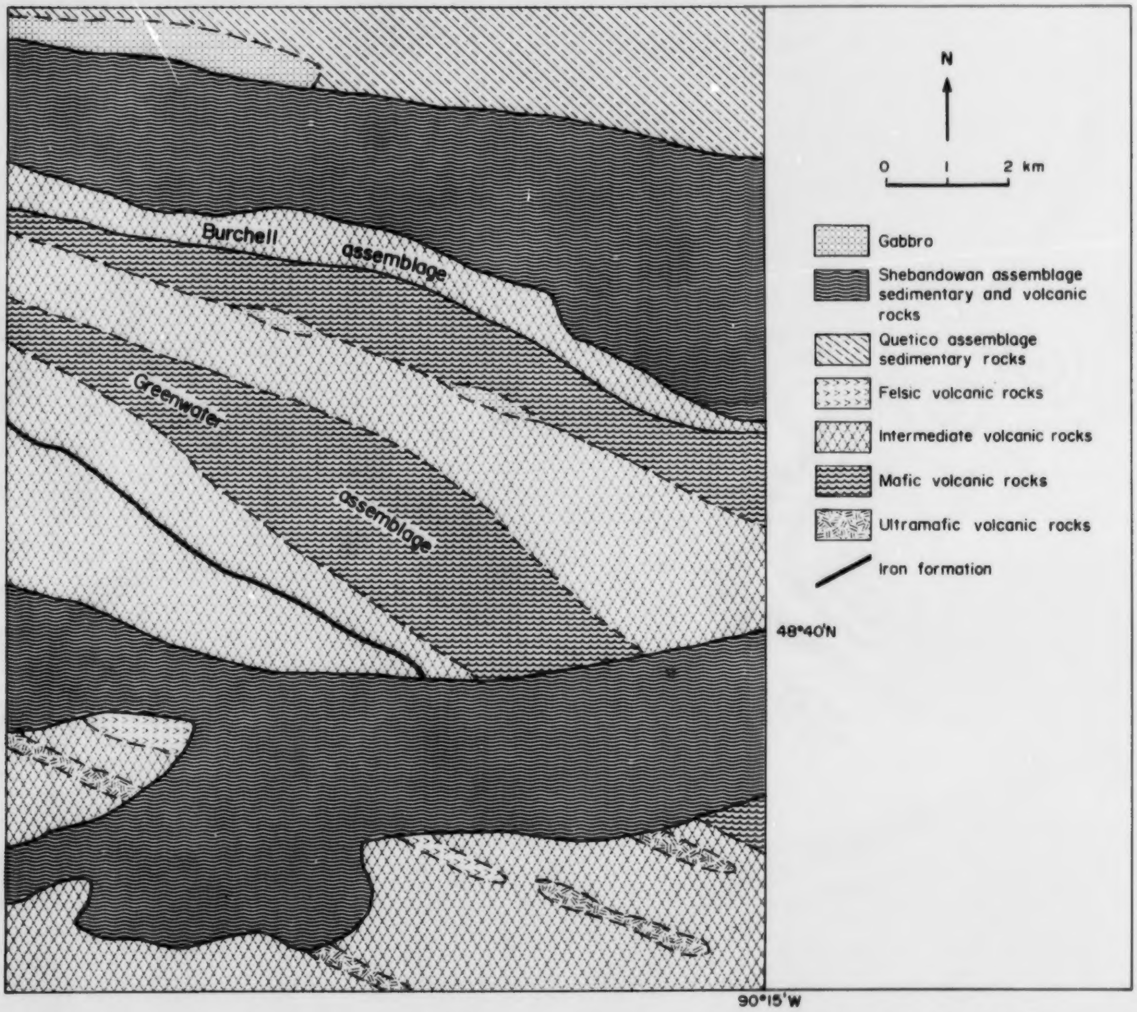

Burchell
assemblage
Greenwater
assemblage
N
0
1
2 km
Gabbro
Shebandowan assemblage sedimentary and volcanic rocks
Quetico assemblage sedimentary rocks
Felsic volcanic rocks
Intermediate volcanic rocks
Mafic volcanic rocks
Ultramafic volcanic rocks
Iron formation
48°40'N
90°15'W

Figure 12.10. An area in Horne Township showing Shebandowan assemblage rocks unconformably overlying rocks of cycle 2 in the Greenwater assemblage (*modified from* Carter 1984).

within the Greenwater assemblage, sills are thinner, finer grained and more intimately interlayered within the volcanic units. Peridotitic compositions are also more typical in the southern half of the belt.

Detailed investigations of the petrology and geochemistry of peridotitic intrusions have been stimulated by an abundance of mafic to ultramafic intrusions, both in this belt and in the nearby Wabigoon Subprovince (Watkinson and Irvine 1964). A copper-nickel ore deposit at the Shebandowan Mine is within one such ultramafic intrusion (Morton 1982).

STRUCTURAL GEOLOGY

The following structural analysis of the Shebandowan–Saganagons greenstone belts relies heavily on work recently carried out by the Ontario Geological Survey in the Shebandowan greenstone belt. The structural geology of the Saganagons belt is described first, followed by a much more extensive analysis of the Shebandowan greenstone belt.

Saganagons Belt and Shebandowan Greenstone Belt

In the Saganagons belt, top direction data from the 2 volcanic assemblages show that they are vertically stacked as oppositely facing homoclinal panels, whereas the Knife Lake sedimentary rocks are folded. Harris (1968) reported that the Saganagons assemblage is slightly overturned to the north, since the volcanic strata young consistently northward and dip steeply southward. Lineations of minerals and grain shapes consistently plunge about 50° southwestward. This contrasts with the northeastward plunging mineral lineations, close to the Saganaga pluton, interpreted to indicate that a contact strain aureole has been imposed by the emplacement of the pluton. The pluton has also imposed a metamorphic aureole on adjacent early volcanic rocks.

The Saganagons belt and external granitoids are cut by faults, some of which show evidence of left-handed transcurrent motion, forming strike-slip duplexes in the supracrustal rocks. An example is the effect of the Knife Lake fault system on the Greenwater assemblage, west of Saganaga Lake, that is shown in Figure 12.8. Similar faults transect and subdivide the Knife Lake Group of northeastern Minnesota into fault-bounded domains, making stratigraphic correlation difficult in the absence of age determinations of the rock units. These faults correspond in age to the late faults transecting the Shebandowan greenstone belt, such as the Crayfish Creek fault. Some of the faults probably have a history of early thrusting or reverse faulting during the juxtaposing and telescoping of supracrustal assemblages. This was probably followed by transcurrent motion and, in some cases, later vertical motion. Only the last increments of motion are preserved by kinematic indicators.

The following description features the structural geology of the Shebandowan greenstone belt; the central part of it was subjected to a regional structural analysis by Stott (1986); the main patterns of regional deformation are described by Stott and Schwerdtner (1981) and Stott and Schnieders (1983). In addition, detailed studies include Morton (1982), Borradaile and Brown (1987) and Kehlenbeck (1988). The following summary is based mainly on Stott (1986). A summary of the characteristics of structural elements of the main deformation events is given in Table 12.4. The regional significance of the structural patterns observed in this belt are discussed later in this chapter.

Earliest Assembly of the Shebandowan Greenstone Belt

In most greenstone belts, the history of crustal assembly of the belt at its earliest stages is not well known. Each of the 2 older assemblages contains several units of felsic volcanic rocks dominantly separated by mafic flows. The generally consistent stratigraphic-facing direction within each assemblage (*see* Figure 12.9) may be interpreted such that each volcanic pile grew progressively by episodic extrusion of mafic to felsic flows and pyroclastic deposits. Alternatively, the cyclicity of volcanism may have a secondary cause, in part reflecting the tectonic interleaving of strata with repetition of all or part of some volcanic cycles. Such

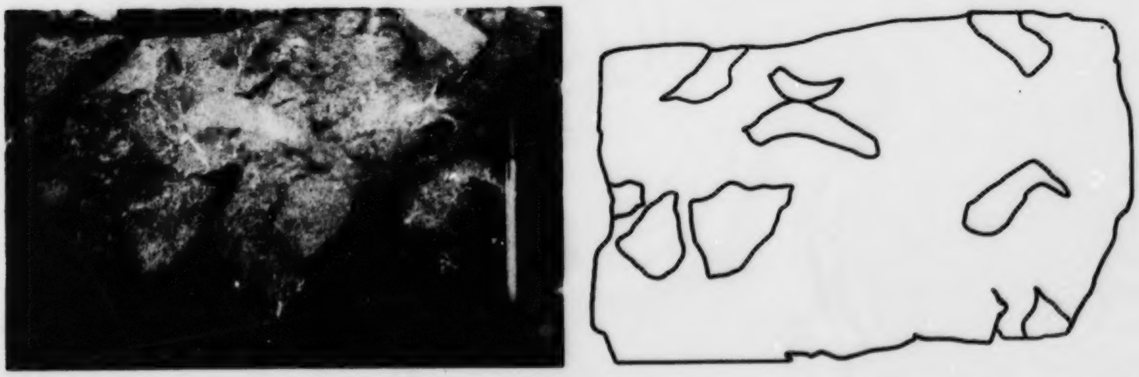


Figure 12.11. A block of latitic volcanic breccia of the Shebandowan group, Timiskaming assemblage, southeast of Lower Shebandowan Lake (pen is 15 cm in length). The sketch shows an enhanced outline of the clasts.

Table 12.4. Mesoscopic structural elements of the Shebandowan belt and their characteristics.

Deformation Phase and Associated Structural Elements	D_1 (F_1, S_1, L_1)	D_2 (F_2, S_2, L_2)	D_3 (F_3, S_3, L_3)
Fold Profile	tight to isoclinal	tight, isoclinal to open in sedimentary rocks and Shebandowan assemblage volcanic rocks	kink folds dominantly Z-folds
Fold Orientation	plunging upright folds with near vertical axial surfaces and moderate to steep rake of fold axes; axes plunge west to southwest	plunging upright folds with near vertical axial surfaces and gentle to steep rake of fold axes; axes plunge east to northeast	plunging upright folds with near vertical axial surfaces and generally steep rake of fold axes; axes plunge variably west, north and east
Fold Distribution	characterizes D_1 strain domains but sparsely observed; best seen in mafic tuffs southwest of the Greenwater Lake granodiorite pluton; probably occur in D_2 domains, but not observed in outcrop	characterizes D_2 domains but sparsely observed except in sedimentary rocks, tuffs and Shebandowan assemblage volcanic rocks	widely observed in north half of belt and in Quetico metasedimentary rocks; most prominent in thinly bedded units
Surface Folded	S_0 bedding	S_0 bedding	S_0 bedding and S_1, S_2 foliations
Axial Surface Foliation	S_1 almost always parallel to bedding; bedding and/or cleavage intersections rarely observed	S_2 parallel to S_1 and bedding; bedding and/or cleavage intersection locally observed in sedimentary rocks and tuffs	S_3 weak and confined to narrow hinge of F_3 folds; commonly a fracture cleavage
Axial Lineations	L_1 mineral and dimensional lineations are parallel to fold axes and plunge west to southwest	L_2 mineral and dimensional lineations are parallel to fold axes and plunge east to northeast	L_3 only locally observed and confined to F_3 fold hinges as crenulations and mineral alignment

interleaving within the assemblages could have occurred during speculative early thrust faulting accompanying the assembly of this belt (Percival and Williams 1989). Precise U-Pb geochronologic studies may reduce uncertainty about the nature and timing of the relationship between these 2 assemblages, which are considered to have been brought together during or prior to D_1 deformation.

D_1 Deformation

The earliest preserved penetrative regional deformation appears to have initially encompassed the entire breadth of the belt, labelled D_1 (Figure 12.12), although earlier thrusting and tectonic interleaving of strata may have occurred. D_1 is characterized by westerly plunging stretching lineations of minerals and mineral-aggregates within a vertical to steeply northward dipping schistosity. Rocks deformed by D_1 alone (Figure 12.13) are less strained than those that enjoyed both D_1 and D_2 deformation events (Figure 12.14).

D_2 Deformation

The second deformation event, D_2, apparently overprinted D_1 in only parts of the belt, most prominently in the north. This produced a pattern of separate domains labelled D_1 and D_2. These domains share a common schistosity but are distinguished by several features, most notably by variation in the plunge of lineations (Figure 12.15), which illustrates the pattern of domains in the central part of the belt. Lineations in the D_1 domains plunge consistently 40° to 60° westward, except near late plutons. In contrast, D_2 lineations plunge to the east but with a greater range of inclinations. Most notable is the rather shallow lineation that characterizes the northwestern part of the belt in the main D_2 domain. This shallow-plunging lineation can be traced across the subprovince boundary into the Quetico Subprovince as a second deformation defined by Sawyer (1983) in wacke sedimentary rocks. Southwest of Burchell Lake, shallow-plunging D_2 lineations steepen and plunge consistently southwestward (Chorlton 1987).

The D_2 event has also produced numerous brittle-ductile shear zones, some of which are hosts to gold mineralization (Stott and Schnieders 1983). Shear zones within the northern D_2 domain display kinematic indications of right-handed displacement east of Burchell Lake and left-handed displacement southwest of Burchell Lake (*see* Figure 12.2). This change in the sense of D_2 shear displacement may be a response to northwesterly directed, regional crustal shortening that is described later in this chapter.

The Shebandowan assemblage shows no record of D_1 deformation; instead, there is evidence of D_2 structural elements, including stretching lineations and folds. These structural elements are transected abruptly by faults bordering this assemblage (Borradaile and Brown 1987; Carter 1988b). Thus, the boundary between D_1 and D_2 domains is at least locally faulted and the evidence indicates that the Shebandowan assemblage was deposited after D_1 deformation, thereby accounting for the unconformity between these rocks and the other 2 assemblages.

Near the boundaries between the Burchell and Greenwater assemblages, some stratigraphic reversals occur because of local D_1 or D_2 folding as observed by Morton (1982), for example, south of the Shebandowan Mine. However, within the Shebandowan assemblage there is considerable D_2 folding of interbedded sedimentary and pyroclastic rocks (Stott 1986; Borradaile and Brown 1987).

D_3 Deformation

Late, steeply plunging kink folds (Figure 12.16) are widespread in the northern half of the belt and are not confined to the D_2 domains. They are most prominent in thinly bedded units of rock exhibiting D_2 shearing and are interpreted to reflect residual west-northwesterly, subhorizontal shortening near the subprovince boundary arising from the regional, crustal shortening that produced the D_2 fabrics in the belt.

Faults

The Shebandowan region is transected by major northeast- and northwest-striking faults. These faults form a complementary set and may have developed in response to regional, subhorizontal, northwesterly directed crustal shortening. Examples include the northeast-striking Knife Lake fault system showing left-handed displacement, and the northwest-striking Crayfish Creek fault with right-handed displacement (*see* Figure 12.2). Later-stage vertical movement is recorded by near-vertical lineations on the fault system at the Wawa–Quetico subprovince boundary. A Quetico Subprovince side-up sense of displacement marks the termination of the structural effects of the Kenoran Orogeny in this region.

Strain and Metamorphic Aureoles Around Late Plutons

Within the Shebandowan greenstone belt, late plutons intruding volcanic rocks are surrounded by both metamorphic and contact strain aureoles resulting from intrusion processes (Schwerdtner et al. 1983; Schwerdtner 1986). The plutons possess a primary igneous fabric formed during the rise and solidification of the plutons. They display no internal metamorphism, apart from limited deuteric alteration, and they have imposed an amphibolite facies contact metamorphic aureole on surrounding units. The process of intrusion has also imposed a strain aureole on the surrounding rocks, seen in the steepening of lineations in the adjacent units (*see* Figure 12.15). In general, the metamorphic aureole is narrower than the strain aureole.

Two examples of intrusions with aureoles have been studied in detail: the crescent-shaped Greenwater Lake granodiorite pluton (Schwerdtner et al. 1983), intruded along the boundary between the greenstone belt and a small tonalite gneiss dome; and the Hood Lake syenite pluton (Schwerdtner 1986), which is located adjacent to a regional fault and a ductile shear zone, and is interpreted to postdate initial formation of these planar weaknesses. The Greenwater Lake granodiorite pluton produced a strain aureole of variable width within the surrounding host rocks. The strain

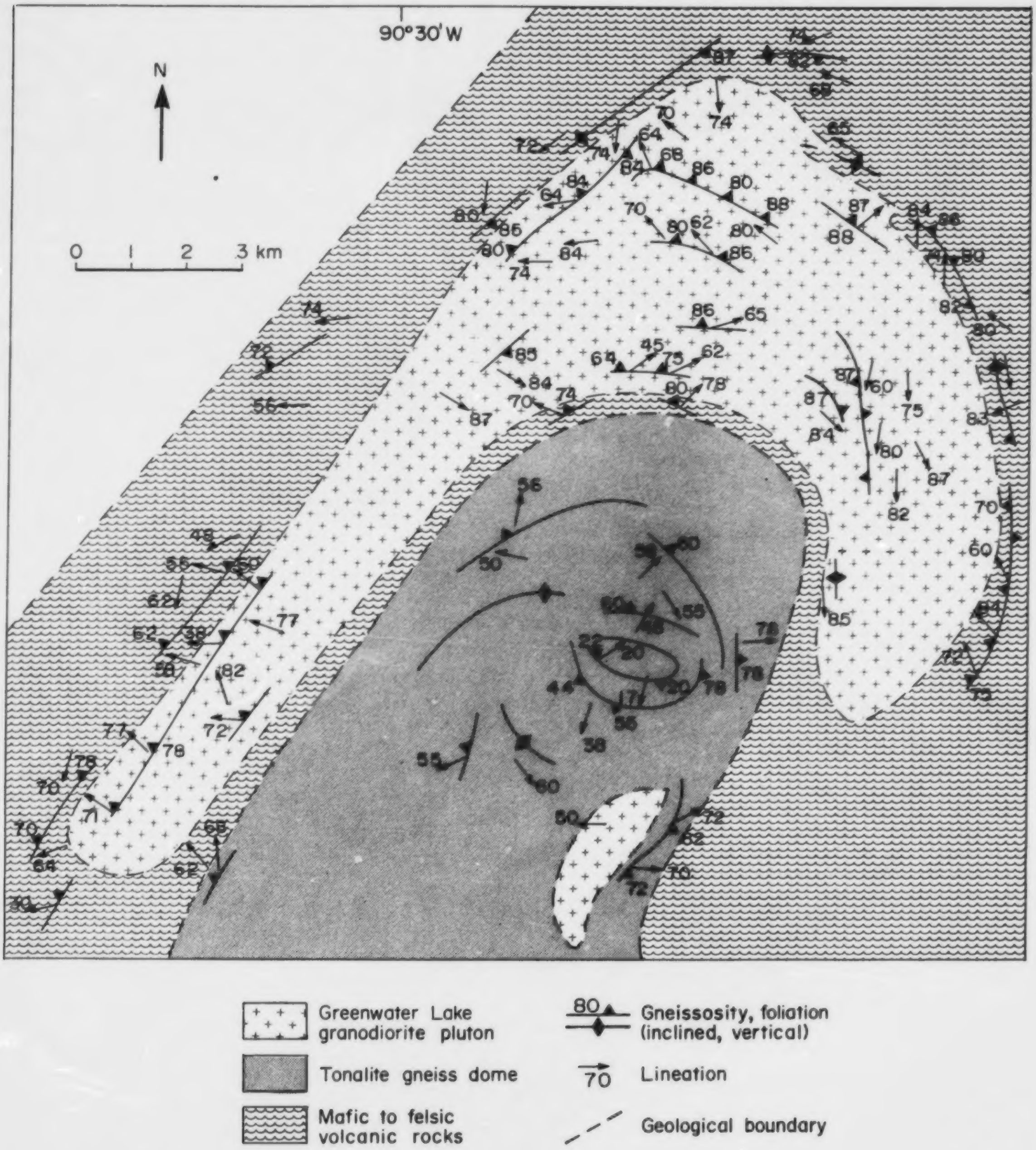

90°30'W
N
0 1 2 3 km
Greenwater Lake granodiorite pluton
Tonalite gneiss dome
Mafic to felsic volcanic rocks
80 Gneissosity, foliation (inclined, vertical)
70 Lineation
Geological boundary

Figure 12.12. Structural elements of the Greenwater Lake granodiorite pluton and its envelope. Note the shallowly dipping gneissosity in the centre of the small tonalite gneiss dome south of the pluton. The pluton appears to have been injected along a previously curviplanar interface between the greenstone belt and a gneiss dome, leaving a thin screen of amphibolite separating the pluton from the gneiss (*modified from* Schwerdtner et al. 1983).

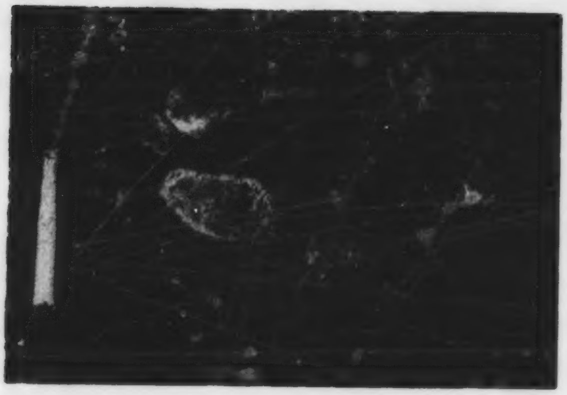


Figure 12.13. Very weakly D_1-deformed basaltic pillow breccia containing well preserved angular hyaloclastite shards in the matrix. Located south of Lower Shebandowan Lake.

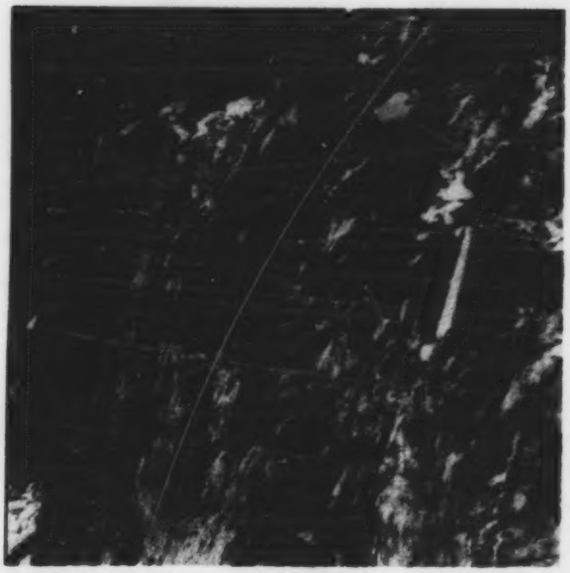


Figure 12.14. Typical fissile, altered volcanic rock in a D_2 shear zone southeast of Burchell Lake.

aureole ranges up to approximately 3 km wide in the volcanic rocks on its eastern margin. The effect of this aureole imposed by the emplacement of the pluton can be seen in the steepening of stretching lineations near the pluton (*see* Figure 12.12).

Other late plutons generally deflect the volcanic stratification and trajectories of regional schistosity and stretching lineations without imposing a significant penetrative change in the inclination of the lineations. These include the Burchell pluton, which has deflected the D_2-strained volcanic rocks around it and thus postdates the penetrative effects of the D_2 event.

METAMORPHISM

A summary of typical metamorphic mineral assemblages in the Shebandowan greenstone belt is given in Table 12.5. Metamorphic grades in this belt range from lower greenschist to lower amphibolite. Greenschist grade mineral assemblages are found throughout the belt, except in areas close to late tectonic to posttectonic felsic plutons where amphibolite grade mineral assemblages are recorded. Examples of contact metamorphism are seen near the Hood Lake syenite pluton where mafic rocks contain poikiloblastic hornblende; and near the eastern margin of the Greenwater Lake granodiorite pluton where amphibolitic basalt grades eastward, away from the intrusion, into greenschist grade assemblages.

Extensive hydrothermal alteration in shear zones is observed in the northern half of the belt, especially north and west of Upper Shebandowan Lake, resulting in carbonatization and sericitization of metamorphic mineral assemblages. At least some of this alteration is intimately associated with D_2 deformation, conjugate faulting and the accompanying gold mineralization (Stott and Schnieders 1983; Chorlton 1987).

GRANITOID ROCKS

Recent work has subdivided the "sea of pink granite" on the maps of the Shebandowan region (Percival 1983; Percival and Stern 1984; Percival et al. 1985). The southern part of the Northern Light–Perching Gull Lakes (NLPG) batholithic complex (*see* Figure 12.2) has been the subject of structural (Grout 1936; Goldich et al. 1972; Davidson 1980) and petrogenetic studies (Hanson 1972; Hanson and Goldich 1972; Stern et al. 1989). Within the Shebandowan greenstone belt, detailed studies have been made on the structure of several plutons and their enveloping host rocks (Schwerdtner et al. 1983; Schwerdtner 1986; Stott 1986). Plutons within the Shebandowan greenstone belt have been useful as relative and absolute time markers (Corfu and Stott 1986) in determining the sequence of tectonic events described below. Within the belt, limited petrogenetic study of the plutons has been made (Smith and Williams 1980).

The oldest recognized pluton within the belt is the Shebandowan Lake pluton, which has been dated at 2696 Ma (Corfu and Stott 1986). This lozenge-shaped intrusion is composed of medium-grained trondhjemite. It is spatially associated with a set of quartz-feldspar porphyry stocks that intrude the surrounding supracrustal rocks. The pluton appears to have intruded prior to D_2 deformation and during, or subsequent to, D_1 deformation since it displays no contact strain aureole upon D_1- or D_2-deformed supracrustal rocks. As well, the southern half of the pluton is notably metamorphosed and D_2 deformed and lies within a D_2 domain. Petrographic descriptions of this pluton, and nearby porphyry stocks, are given by Chorlton (1987).

Rocks within the NLPG batholithic complex range from early phases, consisting of tonalitic gneiss and interleaved supracrustal amphibolite lenses, to posttectonic granite. The pattern of intrusions within this batholithic complex typifies that of many other batholithic complexes

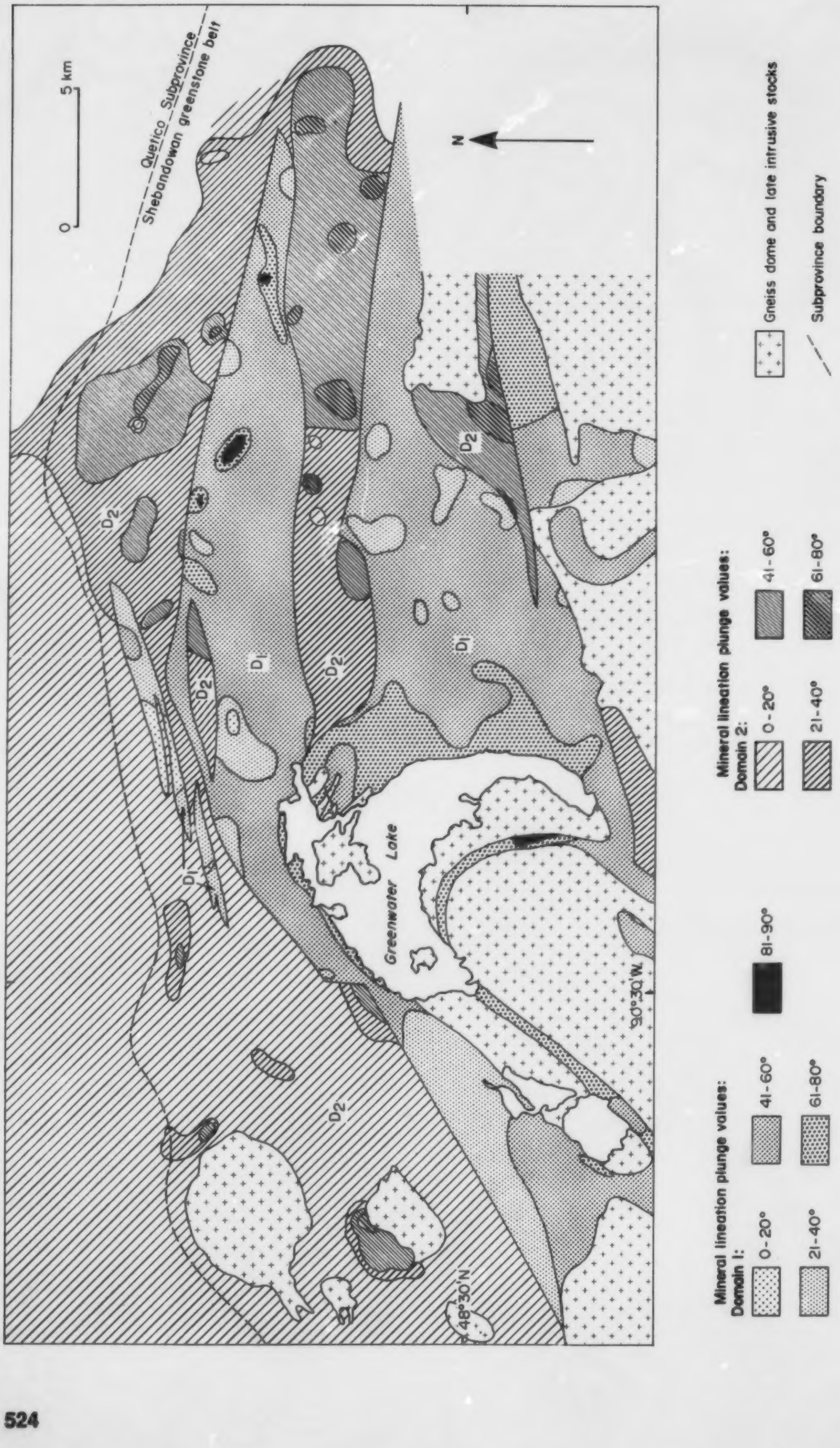

0
5 km
Quetico Subprovince
Shebandowan greenstone belt
N
D2
D1
D2
D2
D1
D2
D1
D2
Greenwater Lake
90°30'W
48°30'N
Mineral lineation plunge values:
Domain 1:
0-20°
21-40°
41-60°
61-80°
81-90°
Mineral lineation plunge values:
Domain 2:
0-20°
21-40°
41-60°
61-80°
Gneiss dome and late intrusive stocks
Subprovince boundary

Figure 12.15. Contoured values of plunges of stretching lineations in the central part of the Shebandowan greenstone belt. Lineations plunge westward in D_1 domains and eastward in D_2 domains. Note the relatively limited range in the plunge inclination of the lineations in the D_1 domain in contrast to those in D_2 domains. The shallowest plunge values are recorded in the northwestern part of the belt and into the Quetico Subprovince. Regional lineations steepen near late plutons (*modified from* Stott 1986).

Figure 12.16. Typical steeply plunging D$_3$ kink folds. Located north of Greenwater Lake and south of Upper Shebandowan Lake.

in the Superior Province; for example, the Uchi Subprovince (*see* Stott, this volume). The core of the complex is dominated by tonalitic gneiss that is typically separated from greenstone belt rocks by crescentic to planar, sheet-like plutons of massive to foliated tonalite, granodiorite or granite. These late plutons either stope their way into place, leaving virtually no penetrative strain effects on the surrounding host rock, or they forcefully intrude as diapirs or sheets. These diapirs or sheets might wedge their way into place along pre-existing faults or curviplanar boundaries between contrasting rock types. This would produce contact strain and metamorphic aureoles upon the enveloping host rocks. An example of a crescentic sheet intrusion with a strain aureole is the Greenwater Lake granodiorite pluton (Schwerdtner et al. 1983; Stott 1986) illustrated in Figure 12.12. The Saganaga pluton exhibits internal primary structural elements such as schistosity, lineations of minerals and quartz aggregates. It is another example of a pluton superimposing a contact strain aureole on the adjacent Saganagons belt (Davidson 1980).

The youngest plutons in the NLPG batholithic complex, and within the Shebandowan greenstone belt, are of the diorite-monzonite-syenite Perching Gull suite. They are interpreted to be similar in age to the Burchell pluton, which is dated at 2684 Ma (Corfu and Stott 1986). These include (*see* Figure 12.2) the Greenwater Lake granodiorite pluton (Stott 1986), Hood Lake syenite pluton (Schwerdtner 1986), the Perching Gull pluton and Icarus pluton (Stern et al. 1989). These intrusions are late tectonic to posttectonic and are characterized by having hornblende-bearing phases and clotty hornblendite aggregates that form stretching lineations (Figure 12.17). The intrusions also contain diorite, monzodiorite or trachyandesite members of the "sanukitoid" suite (Shirey and Hanson 1984), and have rocks with chemical characteristics consistent with derivation by hydrous partial melting of large ion lithophile element-enriched mantle peridotite (Stern et al. 1989).

MINERAL DEPOSITS

Owing to the relative accessibility of the Shebandowan greenstone belt, it has been subject to prospecting interest since at least the beginning of this century. Mineral exploration resulted in the discovery of several different types of deposits and occurrences of base and precious metal mineralization. Several examples are noted here and are depicted on Figure 12.8. Early prospectors discovered iron ore deposits near the Mokomon fault and traced several iron formation units along the belt. In 1913, a nickel-copper sulphide deposit, hosted in a komatiitic flow unit (Lavigne, Aubut and Scott 1990), was discovered on the south shore of Lower Shebandowan Lake, and developed in the early 1970s as the Shebandowan Mine.

The Shebandowan Mine typifies the relatively long time lapse between discovery and the major period of development for several deposits in this belt. One other example was the discovery of copper in the 1870s at the site of the North Coldstream Mine on the east shore of Burchell Lake. It had a sporadic history of development, reaching its peak in the late 1950s and 1960s. Another example of the long time lapse was the discovery of gold in 1871 , southwest of Burchell Lake, with the main periods of development being in the 1930s. This discovery culminated in the Huronian Mine, and in the present day, with the recent development of a copper-gold deposit in that area (Lavigne, Aubut and Scott 1990).

There has been exploration for volcanic massive sulphide (copper-zinc) deposits, exemplified by the Vanguard prospect at the top of cycle 1 of the Burchell assemblage, north of Upper Shebandowan Lake (Hodgkinson 1968; Lavigne, Scott and Sarvas 1990) but with less success than the search for gold prospects in recent years.

Gold deposits are concentrated in the northern half of the Shebandowan greenstone belt (Stott and Schnieders 1983) and locally in the southern part of the belt in the vicinity of Gold Creek (Chorlton 1987). Stott and Schnieders (1983) proposed that gold mineralization within much of this belt is related to the penetrative D$_2$ deformation event in the northern part of the belt, especially in the vicinity, and southwest, of Burchell Lake. The distribution of gold deposits and occurrences illustrated in Figure 12.8 shows a close spatial correlation with the distribution of D$_2$ deformation

Figure 12.17. Steeply plunging primary, magmatic lineations of hornblende and hornblendite clots in the Greenwater Lake granodiorite pluton; one of several late tectonic to posttectonic plutons of the sanukitoid suite (Stern et al. 1989).

Table 12.5. Metamorphic mineral assemblages of the Shebandowan greenstone belt.

Area; References	Rock Type	Observed Assemblage
Burchell Lake; Giblin 1964	**Burchell assemblage**	
	felsic to intermediate volcanic rocks	quartz, albite, white mica, chlorite, carbonate;
	mafic volcanic rocks	chlorite, hornblende, tremolite-actinolite, carbonate, albite, epidote ± quartz; hornblende is largely replaced by chlorite (typical of D_2 domain); narrow chlorite schist bands with carbonate, white mica, albite and quartz
	gabbro (north of Skimpole Lake, east of Burchell Lake)	hornblende, saussuritized plagioclase, quartz, carbonate, epidote, chlorite, apatite, magnetite, pyrite, pyrrhotite
	Greenwater assemblage	
	banded tuff (northwest of Grouse Lake) and tuff in Quetico sedimentary rocks	quartz and minor plagioclase alternating with biotite, minor magnetite and pyrite
	iron formation (north of Squeers Lake)	magnetite, quartz, amphibole ± carbonate ± epidote
	gabbro (north side of Hood Lake syenite pluton)	hornblende, labradorite, minor magnetite, very minor quartz; hornblende is poikiloblastic, reflects contact metamorphism
Kashabowie and, Shebandowan lakes and east of Greenwater Lake; Hodgkinson 1968	**Burchell and Greenwater assemblages**	
	felsic to intermediate volcanic rocks	sericitic schist matrix
	mafic volcanic rocks	chlorite; pale amphibole in southeast and central area, hornblende in north; medium grained amphibolite southeast of Tinto Lake fault
	gabbro	plagioclase, amphibole, chlorite, retrograded locally; local chloritic schist zones
	peridotite	igneous assemblages retrograded; chlorite, secondary amphibole
Lower Shebandowan Lake; Morin 1973	**Burchell assemblage**	
	felsic volcanic rocks	sericitized
	mafic volcanic rocks	greenschist facies: shredded fine-grained chlorite, hornblende, sericite, saussurite, carbonate, albite, coarse-grained feldspar
	gabbro	chlorite, epidote, hornblende, albite
	peridotite (south of Lower Shebandowan Lake)	serpentine, minor talc, magnetite, carbonate
	Shebandowan assemblage	
	argillite	quartz, feldspar, mica
	arkose	microcline, quartz, chlorite, carbonate
south and east of Lower Shebandowan Lake; Shegelski 1980	**Greenwater assemblage**	
	basalt	greenschist facies: sodic plagioclase, chlorite, hornblende, epidote, sericite, saussurite, quartz
	Shebandowan assemblage	
	sedimentary rocks	greenschist facies: hornblende, potassium feldspar, plagioclase, chlorite, sericite, quartz and epidote

Table 12.5. Metamorphic mineral assemblages of the Shebandowan greenstone belt.

Area; References	Rock Type	Observed Assemblage
north of Crayfish Creek fault on Discovery Point; Morton 1982	**Burchell assemblage,** basalt	upper greenschist facies: actinolite, epidote, quartz; actinolite, epidote; fine-grained actinolite, microlites of plagioclase
south of Lower Shebandowan Lake; Morton 1982	basalt	lower greenschist facies: chlorite, quartz, carbonate, plagioclase ($<An_{30}$); chlorite, quartz ±carbonate, epidote, actinolite; chlorite, quartz, epidote, actinolite
		upper greenschist facies: epidote, actinolite ±quartz
southeast of Tinto Lake fault; Morton 1982	**Shebandowan assemblage** metasedimentary rocks	quartz, actinolite, plagioclase
Blackwell, Laurie, Goldie, Horne, Forbes, Conmee townships; Carter 1990a, 1990b, 1990c	**Burchell assemblage** komatiitic volcanic rocks mafic rocks	saussuritized plagioclase, tremolite, actinolite
	ultramafic rocks	chrysotile, carbonate, chlorite, magnetite
	Greenwater assemblage komatiitic volcanic rocks ultramafic rocks	chlorite, serpentinite, tremolite
	mafic rocks	saussurite, quartz, chlorite
	tholeiitic volcanic rocks mafic rocks intermediate rocks	sericitized plagioclase, chlorite, epidote, carbonate; albite, actinolite, biotite, saussuritized plagioclase, chlorite, carbonate; plagioclase, quartz, chlorite, epidote; sericite, albite, chlorite
	calc-alkalic volcanic rocks mafic rocks	saussuritized plagioclase, quartz, carbonate; saussuritized plagioclase, quartz, clinozoisite, chlorite, carbonate
	Shebandowan assemblage alkalic volcanic rocks mafic rocks	sericite, chlorite, carbonate
	intermediate rocks	sericitized plagioclase, epidote, actinolite, chlorite

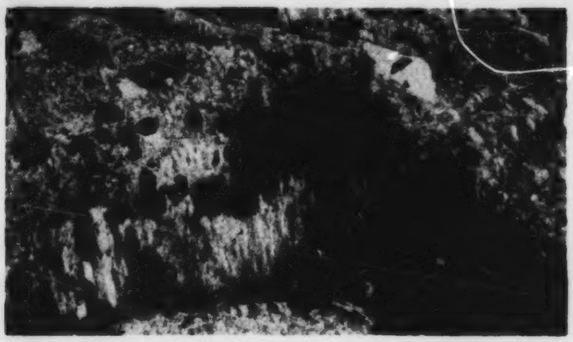


Figure 12.18. Steeply plunging lineations in a dacitic lapilli tuff of upper cycle 2–Greenwater assemblage within the contact strain aureole of the Kekekuab Lake pluton.

domains. This was confirmed (Chorlton 1987) within the northern D_2 domain, in a detailed study of several deposits near Highway 11 and at the former Huronian Mine. Chorlton noted, especially, the role of brittle-ductile shear zones bearing carbonate-sericite alteration and auriferous quartz veins.

The Gold Creek area in the southern part of the Shebandowan greenstone belt displays a different relationship of gold mineralization to deformation. According to Chorlton (1987, p.228), gold mineralization is associated with sericitic and pyritic alteration zones contained in the most highly fractured and brecciated areas of the felsic to intermediate volcanic rocks in the Greenwater assemblage—upper cycle 2.

It is suspected, from structural and relative timing relationships, that fracturing and mineralization is related to the emplacement of the late tectonic Kekekuab Lake pluton on the southern margin of the belt (*see* Figure 12.8). This pluton has produced a broad contact strain aureole marked by strongly developed subvertical stretching lineations (Stott 1986), as shown on Figure 12.18. The broad contact strain aureole encompasses the mineralized zones of the Gold Creek area. This mineralization, if confirmed with additional studies, would exemplify a deposit-type (e.g., Dona Lake Mine, Uchi Subprovince) that is related to late pluton-induced deformation and hydrothermal alteration. Such a deposit differs in origin from those that occur within the northern part of the Shebandowan greenstone belt.

The orientation of ore shoots at several deposits in this belt corresponds to the direction of subsequent tectonic stretching within the D_2 domains in the northern half of the belt. The plunge direction or shape of the ore in the Huronian gold mine, the North Coldstream copper mine and the Shebandowan nickel-copper mine is parallel in each case with the characteristic direction of ductile stretching within the host rocks. This is evident by comparing the mine sections with the orientation of D_2 stretching lineations in the host rocks of each mineral deposit (Stott 1986; Chorlton 1987). For example, the nickel-copper ore of the Shebandowan Mine forms a ribbon-like mass that plunges eastward, parallel to the local D_2 stretching lineations (*see* Figure 12.12).

TECTONIC SEQUENCE OF EVENTS

A general history of tectonic events proposed for the Shebandowan greenstone belt by Stott and Schnieders (1983) and Stott (1986) is summarized in Table 12.6. It incorporates age determinations for several rock units in the Shebandowan greenstone belt (Corfu and Stott 1986). The geochronologically defined events are illustrated in Figure 12.19. The Shebandowan Lake pluton may have intruded during or subsequent to D_1 deformation; part of this intrusion is affected by D_2 deformation. The latitic pyroclastic breccia of the Shebandowan assemblage was deformed by D_2 and provides a minimum age for the initiation of D_2; whereas the Burchell pluton deflects D_2 deformation in the volcanic rocks around it and provides a minimum age for D_2 penetrative activity. D_2 is postdated by a suite of plutons, most of which are members of the "sanukitoid suite" of Stern et al. (1989). They are compositionally and temporally correlative with the Ottertail Lake pluton straddling the Quetico Fault near the Wabigoon–Quetico subprovince boundary and provide an important age constraint on the timing of regional transpressive (D_2) deformation across this southern region of the Superior Province.

The volcanic assemblages of the Shebandowan and Saganagons belts are lithologically similar to those of modern island-arcs and marginal basins, but their original depositional environments remain open to investigation. Their apparent tectonic juxtaposition during the assembly of the Shebandowan greenstone belt occurred between approximately 2733 Ma (a minimum age for volcanism within cycle 2 of the Burchell assemblage) and 2689 Ma (the age of latitic pyroclastic breccia in the unconformably overlying Shebandowan assemblage).

There is some uncertainty in the age of D_1 deformation and of its role in the history of the Shebandowan greenstone belt. The Shebandowan Lake pluton (2696 Ma) is unaffected by D_2 deformation in its central core, and shows evidence of primary magmatic textures and quartz aggregate lineations that parallel the D_1 lineations in the volcanic rocks. It could, therefore, be argued that the pluton intruded during, or subsequent to, D_1 deformation but before D_2 deformation. Alternatively, the D_1 deformation is related to and followed closely by D_2 deformation.

D_2 deformation affects a pyroclastic breccia unit of the Shebandowan assemblage, dated at about 2689 Ma, but predates the Burchell pluton dated at about 2684 Ma. The age of deformation is therefore closely bracketed.

The pattern of deformation in the belt is consistent with a regional northwesterly directed subhorizontal compression (transpression). This culminates in D_2 deformation, and the termination in transcurrent fault movements, kink folding near the subprovince boundary, and vertical uplift of the Quetico Subprovince relative to the Shebandowan greenstone belt. The timing of magmatic and structural events compares closely with similar activity across the Wabigoon–Quetico subprovince boundary described by Davis et al. (1989) and is schematically illustrated in Figure 12.20.

LATE NEOARCHEAN TO PROTEROZOIC ACTIVITY

DIABASE DIKES

Diabase dikes cutting greenstone belts and the accompanying granitoids are described in detail by Osmani (this volume). The generally northerly trending Matachewan dikes were intruded during late Neoarchean–Paleoproterozoic extension, and represent a continuance of a stress regime that prevailed during the coalescence of the Wawa, Quetico and Wabigoon subprovinces. The Wawa Subprovince is cut by the 1 billion-year-old (Gittins et al. 1967) Firesand alkalic complex and the 1.1 billion-year-old Port Coldwell alkalic complex (Heaman and Machado 1987). Details of these rocks are described by Sage (this volume).

FAULTING

Recent research (e.g., Peterman and Day 1989; Kamineni et al. 1990) has revealed a protracted history of events involving motion on major strike-slip faults that in part form subprovince boundaries (e.g., the Quetico Fault). By analogy, the authors infer that brittle faulting continued to be active after coalescence of the subprovince with its tectonic neighbours. This occurred within 1) the Wawa Subprovince (e.g., the Lake Superior–Hemlo fault zone); 2) the boundaries of the Wawa Subprovince (e.g., the Kapuskasing Structural Zone in the east, Percival 1990); and 3) the strike-slip fault systems that form the northern and western edges of the Shebandowan greenstone belt.

Fault patterns and senses of fault displacement, as described for the Shebandowan greenstone belt, are similar to those elsewhere in the subprovince. The authors consider that the cause of faulting was continued subprovince convergence.

DISCUSSION

The following sections briefly discuss a number of aspects of Archean geology as exemplified by the Wawa Subprovince.

CORRELATION OF ASSEMBLAGES BETWEEN AND WITHIN BELTS

In the eastern part of the Wawa Subprovince, 2 groups of assemblages appear to have consistent relationships. The Wawa and Hemlo assemblages structurally, or unconformably, underlie the Catfish and Heron Bay assemblages. The authors' correlation of the Wawa assemblage with the Hemlo assemblage, and the Catfish assemblage with the Heron Bay assemblage, rests on similar lithologic, geochemical and geochronologic characteristics. Despite these similarities it is not known whether these assemblages were ever contiguous.

SUBSTRATE DURING FORMATION OF GREENSTONE BELTS

Lead isotope data from veins within the oldest supracrustal rocks of the Michipicoten greenstone belt (Thorpe 1987) yield 2.88 billion-year-old model ages with a *mu* of 10. This is consistent with the presence of evolved sialic crust older than the 2.89 billion-year-old Hawk assemblage volcanic rocks. The description of the Hawk assemblage notes that there is no vestige of a basement to these old supracrustal rocks. The authors infer tectonic similarity with younger episodes of felsic volcanism; thus, the Hawk volcanic rocks are most likely to have been derived from the melting of existing sialic crust.

It is also noted that in the Michipicoten belt, the structural relationships of its oldest assemblage, the Hawk assemblage, are such that it lies along the northwestern edge of the Whitefish Lake batholith. This relationship is similar to that of the relatively old Pacaud assemblage around the eastern margin of the Round Lake batholith in the Abitibi Subprovince (*see* Jackson and Fyon, this volume). Perhaps the emplacement of large batholiths brings older, once deeply buried supracrustal assemblages to the surface. Alternatively, perhaps the batholiths represent, in part, a remobilized substrate to the early assemblages. This suggestion has yet to be tested with isotope studies.

INFERRED TECTONIC SETTINGS OF ASSEMBLAGES

Subdivision of greenstone belts into assemblages is one brought about by reason of: 1) mapped complexities of structure and stratigraphy; and 2) precise geochronologic data. In accord with the work of Thurston and Chivers (1990), the authors have recognized 4 differing types of supracrustal assemblage, representing: mafic plain volcanism (oceanic?); tholeiitic and calc-alkalic volcanism (island arc?); wacke-iron formation association (fore-arc, back-arc basin?); and pull-apart or strike-slip basin environments. Tables 12.2 and 12.3 indicate interpretations of the tectonic environment of development for most of the assemblages. A brief discussion of the tectonic significance of these environments on a subprovince and "Superior Province" scale is discussed in a later chapter (*see* Williams, Stott, Thurston et al., this volume).

STRUCTURAL STYLES IN THE WAWA SUBPROVINCE

A record of an early thrusting or recumbent folding event is emerging from the well-preserved rocks within the Michipicoten greenstone belt. This early history is only apparent in broad greenstone belts where the volume of rocks has been sufficient to preserve relicts which have withstood deformation and plutonism. In the higher-grade, tonalitic gneiss terranes located along the northern edge of the subprovince, tonalite sheets were emplaced along shearing surfaces at the margins of, and engulfing, relatively thin supracrustal units. Subsequent to the formation of recumbent structures, upright belt-parallel folds, often with steep plunges, reworked the early flat-lying folds and slides. Later, strike-slip deformation along ductile shears and faults, caused the formation of pull-apart basins such as the Shebandowan and Knife Lake assemblages, which were then deformed by continued subprovince convergence or transpression(?).

The cratonization of the Superior Province during the late stage of the Kenoran Orogeny was accomplished under a northwest-directed horizontal compression, which was at an oblique angle to the westerly trending subprovinces (Schwerdtner et al. 1979; Park 1981). This compression has produced a pattern across the Superior Province of westerly and northwesterly striking dextral transcurrent faults, and less common, northeasterly striking sinistral faults. The compression has also produced broad zones of transpression centred on the metasedimentary subprovinces and including adjacent margins of more rigid greenstone-granite subprovinces. Transpressive deformation occurs through oblique convergence of 2 crustal blocks or lithospheric plates (Harland 1971). The collision of obliquely converging lithospheric plates has been commonly observed in Phanerozoic rocks such as those found in New Zealand (Norris et al. 1990), and a similar scenario is postulated during late Archean convergent tectonics in the Superior Province.

In northwestern Ontario, the zones of transpression exhibit a combination of both crustal shortening and simple shear. This results in a concomitant development of folds and wrench-fault zones. The folds are commonly at a small angle to the strike of the transpression zone (Sanderson and Marchini 1984). Other features typical of transpression zones have been described in the Quetico Subprovince by Borradaile et al. (1988). They noted that the axial planes of upright folds within the northern part of the Quetico Subprovince, near the Wabigoon subprovince boundary, strike obliquely in a counterclockwise sense to the Quetico Fault along that boundary. This orientation of folds is consistent with a northwesterly directed dextral transpression.

Similarly, in the southern part of the Quetico Subprovince, Sawyer (1983) observed late, upright folds to be transected by a cleavage striking slightly counterclockwise to the axial plane of the folds. This geometric pattern is consistent with transecting cleavage-fold relations that Soper (1986) and Blewett and Pickering (1988) interpreted to result from noncoaxial strains in a transpression zone. Caution in the application of this relationship to regional kinematics has been raised by Lafrance et al. (1989). Nevertheless, in the Shebandowan area, one can observe the continuation of the shallow plunge of dimensional and mineral lineations, that are associated with the upright folds in the Quetico Subprovince, into the northern half of the Shebandowan greenstone belt (*see* Figure 12.15).

The Shebandowan greenstone belt forms an arcuate convex boundary against the Quetico Subprovince. The most pronounced curvature of the belt occurs along a northwesterly striking hinge-line near the Burchell pluton (*see* Figure 12.8). The intensity of flattening, and the prominence of concomitant ductile shear zones, is most evident in this part of the belt. The deformation is characterized by subhorizontal mineral and dimensional lineations that progressively steepen away from this area along the length of the belt (Stott 1986; Chorlton 1987). In addition to this symmetric disposition of lineations, there is a corresponding preponderance of dextral strike-slip shear zones in the eastern limb of the belt, and sinistral shear zones southwest of the convex bend in the belt. This symmetry pattern of D_2 deformation centred on the Burchell Lake area is consistent with a northwesterly direction of principal D_2 shortening, and is accordingly consistent with oblique convergence of crustal blocks (the Wawa and Wabigoon) during the late stage of the Kenoran Orogeny.

These structures complement other structures observed in the Vermilion belt of Minnesota and in the adjacent Quetico gneisses by Hudleston et al. (1988) and Bauer (1985), respectively. The Vermilion belt is a discontinuous extension of the Shebandowan greenstone belt. Hudleston et al. (1988) and Bauer (1985) attributed the asymmetric Z-folds and dextral transcurrent shear zones to transpressive D_2 deformation during oblique convergence of 2 rigid lithospheric blocks.

STRUCTURAL SUMMARY

Descriptions of the structural development of each greenstone belt reveal some uniformity of structural style and timing within the subprovince. This section summarizes the structural data and interprets from them the patterns and significance of structural development.

The repetitive nature of some greenstone successions (e.g., the Catfish assemblage in the Michipicoten greenstone belt; the Burchell and Greenwater assemblages in the Shebandowan greenstone belt) are interpreted as examples of, or candidates for, thrust-repetition of strata. In support of this style of deformation within the Michipicoten greenstone belt, Arias and Helmstaedt (1990) described the occurrence of a recumbent fold that later was subjected to upright folding and imbrication. In the Manitouwadge–Hornepayne belt, Williams and Breaks (1990) described the occurrence of flat-lying thrusts and tonalite sheets that were formed during the earliest stages of deformation. The authors conclude that the horizontal tectonic style, as seen in these examples, is symptomatic of the juxtaposition of greenstone assemblages and the emplacement of tonalitic magmas.

A later stage of deformation within greenstone belts produced upright folds, with local imbrication, to form fault stacks, as in the Michipicoten greenstone belt. After folding, there was a period of ductile and brittle shearing in which shear-related folds, shear zones and faults were produced.

DISTRIBUTION OF AGES IN THE WAWA SUBPROVINCE

There is a general pattern of ages of supracrustal rocks emerging from the Wawa Subprovince. Along the northern edge of the subprovince, greenstones of the Burchell Lake, Chapleau and Hemlo–Black River assemblages have furnished ages of around 2.77 to 2.73 Ga. These ages are correlated with the Manitouwadge–Hornepayne, Dayohessarah–Kabinakagami and Schreiber assemblages. Individual assemblages in greenstones are bordered on their southern

Table 12.6. Stratigraph[illegible] and tectonic sequence of events in the Shebandowan greenstone belt.

Structure	Intrusive and/or Extrusive	Criteria
D_3 kink bands in northern half of belt; D_2 cleavage-parallel compression or shearing		time relation to conjugate faults is uncertain; kink bands transect late red felsic dikes which postdate D_2 structures in host rocks; these dikes may relate to posttectonic plutonism in the belt; kink bands may reflect a late shear proximal and subparallel to Quetico–Shebandowan boundary; this layer-parallel shear complicates interpretation of D_2 shear kinematics on foliation surfaces
	posttectonic plutons (Greenwater Lake granodiorite, Kekekuab Lake, Peewatai); associated (?) red felsic stocks and dikes striking oblique to the stratigraphy; wide contact strain aureoles around some plutons	Burchell pluton $2683.7^{+5.8}_{-2.9}$ Ma possesses primary internal fabric that deflects D_2-deformed stratigraphy
conjugate fractures and faults offsetting the stratigraphy; generally northwest-striking dextral and northeast-striking sinistral transcurrent movement		time relation to posttectonic plutons is uncertain; faults represent a late northerly directed compression following D_2 transpression event
D_2 deformation, concentrated in northern half of belt and in Timiskaming-type sedimentary rocks and pyroclastic rocks (Shebandowan assemblage), and in Quetico sedimentary rocks; significant transcurrent shear zones present; this deformation is interpreted to be a regional expression of north- to northwest-directed compression across the Superior Province	major hydrothermal veining during late D_2 deformation, gold-quartz-carbonate veins along D_2 cleavages and late fractures	predates Burchell pluton and postdates Timiskaming-type volcanic and sedimentary rocks and the Shebandowan Lake pluton; D_2 event is bracketed between 2684 and 2689 Ma in the Shebandowan greenstone belt, comparable to the bracket of 2686 to 2692 Ma for the Quetico Fault (Davis et al. 1989); D_2 domains contain lode gold deposits in syn-D_2 hydraulic fractures
	extrusion of Timiskaming–type volcanic rocks, deposition of clastic sedimentary rocks, and iron formation, as the Shebandowan assemblage within linear (graben?) basins; penecontemporaneous erosion of greenstones; deposition of sediments in the Quetico basin to north	latite pyroclastic breccia dated at $2689.3^{+2.5}_{-2.2}$ Ma and associated conglomerate contains a trondhjemite clast dated at $2704.1^{+1.8}_{-1.6}$ Ma; recording significant erosion and linear basin development during and subsequent to the D_1 event
	Shebandowan Lake pluton intruded $2696.1^{+2.4}_{-2.3}$ Ma during or subsequent to D_1 deformation; shape conforms to structural grain of the belt; pluton straddles Burchell and Greenwater assemblages and defines minimum age of their tectonic juxtaposition	pluton possesses primary granitic texture except where affected by D_2 deformation; primary texture indicates pluton postdates D_1 event

Table 12.6. Stratigraphic and tectonic sequence of events in the Shebandowan greenstone belt.

Structure	Intrusive and/or Extrusive	Criteria
D_1 deformation, during or subsequent to the juxtaposition of Burchell and Greenwater assemblages; westerly plunging upright folding of supracrustal rocks across the Shebandowan greenstone belt; record of D_1 is unaffected by D_2 or late-pluton strain aureoles and is preserved only in parts of the southern half of belt; D_1 deformation is attributed to an early phase of regional tectonic shortening culminating in later D_2 deformation	tonalitic diapirism assumed as first stage development of Northern Light-Perching Gull Lakes batholitic complex; subsequent intrusions were emplaced and may span tectonic record to the posttectonic diorite-granite Perching Gull suite (e.g., Perching Gull pluton and Greenwater Lake granodiorite pluton), which include sanukitoid suite rocks	period of erosion during and subsequent to D_1, in which deposits of Timiskaming-type rocks (Shebandowan assemblage) unconformably overlie D_1 deformed Burchell and Greenwater volcanic rocks
	volcanism of the Burchell and Greenwater assemblages; several cycles evident with bimodal suites of tholeiitic basalt and calc-alkalic dacite, rhyolite; and a more evolved suite along the centre of the belt in the Greenwater assemblage of komatiite, tholeiite, and calc-alkalic rocks	stratigraphic zones of intermediate to felsic volcanic rocks infer separate cycles; contrast of regional stratigraphic-younging sense between assemblages; a felsic sill near the northern margin of the Shebandowan greenstone belt, dated at $2733.3^{+2.9}_{-2.5}$ Ma, corresponds in age with felsic volcanism and plutonism in the Wabigoon belt near Mine Centre at about 2730 Ma and provides a minimum age for part of the Burchell assemblage

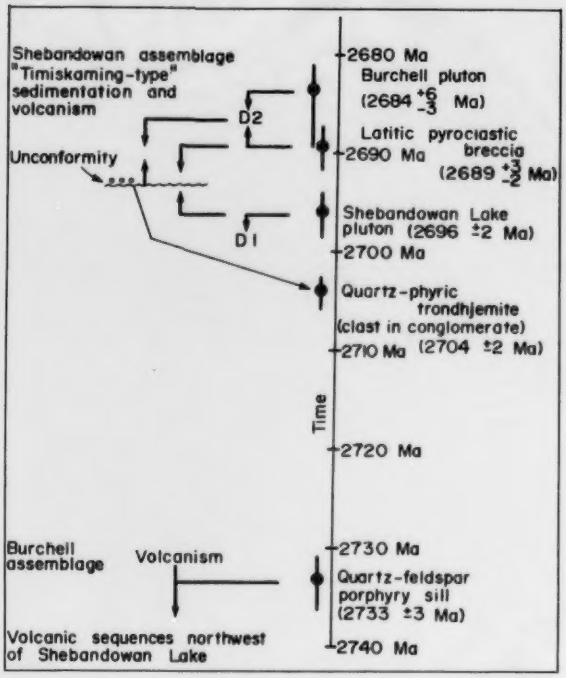

Shebandowan assemblage
"Timiskaming-type" sedimentation and volcanism
Unconformity
D2
D1
Burchell assemblage
Volcanism
Volcanic sequences northwest of Shebandowan Lake
Time
2680 Ma
Burchell pluton
(2684 +6 -3 Ma)
Latitic pyroclastic breccia
2690 Ma
(2689 +3 -2 Ma)
Shebandowan Lake pluton (2696 ±2 Ma)
2700 Ma
Quartz-phyric trondhjemite
(clast in conglomerate)
2710 Ma
(2704 ±2 Ma)
2720 Ma
2730 Ma
Quartz-feldspar porphyry sill
(2733 ±3 Ma)
2740 Ma

Figure 12.19. Sequence of geochronologic events in the tectonic history of the Shebandowan greenstone belt based on Corfu and Stott (1986).

side by high strain zones at Shebandowan and by younger supracrustal rocks formed at around 2.7 Ga at Hemlo.

The oldest rocks in the subprovince, the Hawk assemblage, occur at the margin of a later tonalitic batholith. Perhaps the assemblage has been preferentially exposed there by uplift during, or as a result of, emplacement of the igneous body (cf. the Pacaud assemblage adjacent to the Round Lake batholith; *see* Jackson and Fyon, this volume).

Many greenstone belts contain 2 assemblages of different ages (2.77 and 2.70 Ga), yet have remarkably and distressingly similar lithologic characteristics, thus making distinction difficult.

The spread of ages, determined from the granitoid rocks that form much of the rock between the major greenstone belts, is large. By analogy with Wabigoon Subprovince data (*see* Blackburn et al., this volume), the authors consider that much of the tonalite between greenstone belts is synchronous with the development of the calc-alkalic volcanic rocks in the neighbouring greenstone belts. For example, beside the 2.77 to 2.75 billion-year-old Hemlo (Corfu and Muir 1989a) and Wawa (Turek et al. 1984) assemblages, tonalites have been dated as old as 2.72 Ga. However, other granitoid rocks within these large areas are considerably younger, as young as 2.67 Ga, forming crescentic plutons at the margins of greenstone belts (Turek, Sage and Van Schmus 1990, Dubreuilville pluton; *see* Figure 12.4).

Granitoid rocks from the Kapuskasing Structural Zone have furnished ages younger than 2.67 Ga (Percival 1990), and represent a locally preserved fraction of very young granitoid magmatism. Geochronologic data support the

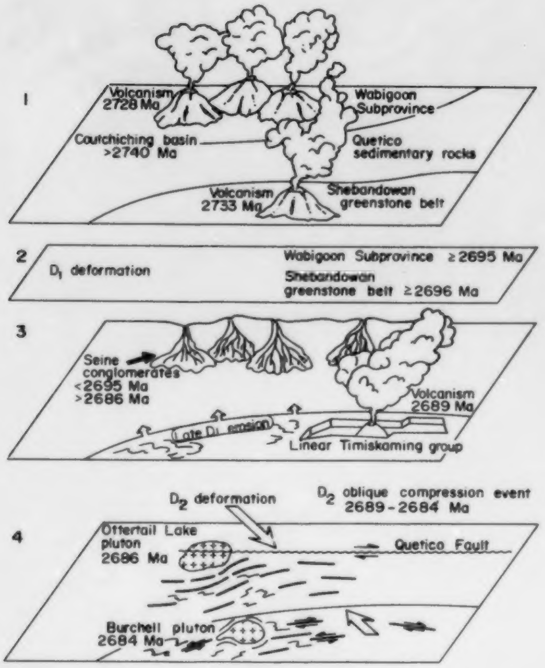

1
Volcanism 2728 Ma
Wabigoon Subprovince
Couchiching basin >2740 Ma
Quetico sedimentary rocks
Volcanism 2733 Ma
Shebandowan greenstone belt
2
D_1 deformation
Wabigoon Subprovince ≥2695 Ma
Shebandowan greenstone belt ≥2696 Ma
3
Seine conglomerates <2695 Ma >2686 Ma
Volcanism 2689 Ma
Late D_1 erosion
Linear Timiskaming group
D_2 deformation
D_2 oblique compression event 2689-2684 Ma
4
Ottertail Lake pluton 2686 Ma
Quetico Fault
Burchell pluton 2684 Ma

Figure 12.20. A 4 stage sequence of events across the Wabigoon–Quetico–Wawa subprovince boundaries. Early thrusting (*see* Williams, this volume) and penetrative deformation (D_1 of Shebandowan) follow episodes of volcanism and sedimentation in both the Wabigoon and Shebandowan greenstone belts (*modified from* Stott 1986; *modified from* Davis et al. 1989).

contention that magmatism continued at a deep tectonic level in granulite facies rocks at a time when magmatism had already ceased at higher tectonic levels. Comparison of late-tectonic magmatic ages between rocks of differing tectonic level seem to indicate that magmatism retreated to deeper tectonic levels with time. In regard to the ages of magmatism within and adjacent to greenstone belts, it is noteworthy that the timing of hydrothermal alteration associated with the Hemlo gold deposits postdates greenstone plutonism but is contemporaneous with magmatism at deep tectonic levels that has been noted by Percival and Card (1985) and Corfu and Muir (1989b). Perhaps the driving force for hydrothermal alteration and mobilization of gold is from a deep source (Fyon et al. 1989).

METAMORPHIC GRADE DISTRIBUTION AND CAUSES

Large changes in regional metamorphic grade are apparent within the Wawa Subprovince. We assert that the changes are, in part, related to the ratio of supracrustal to plutonic rocks. Lowest grades of metamorphism are found within large greenstone belts that contain few internal granitoids (e.g., the Michipicoten and Abitibi greenstone belts), and the highest grades of metamorphism are found within smaller greenstone belts that contain more internal granitoids (e.g., the Manitouwadge greenstone belt). In addition, it appears that some greenstone belts not only have higher ambient metamorphic grades, but themselves contain greater amounts of intrusive granitoid rocks. Grades of metamorphism are low greenschist within the Michipicoten greenstone belt (T = 325° to 450°C; P = 0.2 to 0.3 GPa; Studemeister 1983), rising to amphibolite facies amongst the supracrustal septa and enclaves within tonalitic rocks located northeast of Renabie (Heather and Buck 1988). Metamorphic grades reach almandine-amphibolite facies at Hemlo (T = 455° to 520°C; P = 0.6 GPa; Pan 1990) and at Manitouwadge (T = 750°C; P = 0.45 to 0.6 GPa; Robinson 1979), and approach granulite facies (T inferred at 700°C; P = 0.47 to 0.59 GPa; Percival 1989) at the southern edge of the Quetico Subprovince, adjacent to the Wawa Subprovince in the Manitouwadge–Hornepayne greenstone belt (Williams and Breaks 1989, 1990). We consider that the temperature and pressure variation is a result of tectonic and magmatic thickening processes involving the emplacement of tonalite sheets. The thickening resulted in uplift, which to a large degree controlled subsequent erosion and exposure of metamorphic rocks.

TIMING OF SUBPROVINCE ASSEMBLY

The authors conclude that the timing of assembly of assemblages to form greenstone belts must precede the coalescence of subprovinces to form the Superior Province (Williams 1990). Turek, Sage and Van Schmus (1990) considered that an age of 2677±6.9 Ma from a breccia within the Mishibishu greenstone belt records the youngest volcanic event in the Wawa Subprovince. This age makes it younger than not only the coalescence of assemblages to form the Wawa Subprovince, but also the coalescence of subprovinces to form the Superior Province (Corfu and Stott 1986; Percival and Williams 1989).

The volcanic status of this rock is equivocal, therefore, the youngest regionally developed greenstones within the Wawa Subprovince are those of the Catfish (2698 to 2693 Ma) and Heron Bay (2695 Ma) assemblages, which are slightly younger than the Gamitagama assemblage (2713 Ma). All of these rocks have been intruded by undeformed plutons of granite and granodiorite at about 2688 Ma. Deformation of the greenstones, including structural juxtaposition of contrasting assemblages, is considered to have occurred after approximately 2695 Ma, but prior to the emplacement of undeformed plutons. Also, the assembly of subprovinces into the Superior Province was substantially complete by 2688 Ma. These considerations are in accord with those arrived at for other subprovinces. After the assembly of the Superior Province, there was sporadic plutonism during the time period 2680 to 2650 Ma, during which time, further shearing, hydrothermal alteration and strike-slip faulting were operative.

WAWA–QUETICO SUBPROVINCE RELATIONS

The relations between the Wawa and Quetico subprovinces are discussed fully in Stott et al. (this volume) and were subject of a study by Percival and Williams (1989) and Williams (1990). The Wawa Subprovince structurally underlies the southernmost section of the Quetico accretionary prism. Corfu and Stott (1986) demonstrated that the coalescence of the 2 subprovinces was probably ongoing by 2696 Ma but was substantially complete by 2684 Ma. The contact zone is a northerly dipping zone of highly sheared rocks, locally (e.g., at Manitouwadge) at high grades of metamorphism. Kinematic indicators generally, but not without exception (e.g., at the Shebandowan area), show north-over-south displacement senses.

Stott (1986) mentioned that the northern margin of the Wawa Subprovince displays a saw-tooth pattern. Stott considered that the nonlinearity of the boundary is, in part, original, and represents irregularities in the arc-outline prior to the coalescence with the overriding Quetico Subprovince. The pattern is not unlike a chain of indentors, as envisaged by Molnar and Tapponier (1975), in their hypothesis of slip-line field tectonics for the Himalayas. Kinematic indicators, such as mineral lineations, fold asymmetry and displacement vectors on faults, display contrasting orientations on either side of a promontory, which correspond with high-strain zones.

CHAPLEAU BLOCK RELATIONS

Percival and Card (1983, 1985) considered that the observed variation in rock type, structure, metamorphism, gravity and aeromagnetic expression, were best explained by the hypothesis that the Chapleau block represents an uplifted midcrustal equivalent to the Michipicoten greenstone belt. Percival et al. (1989) found additional support for this hypothesis in reflection seismic profiles which trace the structures of the Chapleau block westward to midcrustal depths. The authors, however, are impressed by the lithologic and structural similarities between rocks exposed along the northern edge of the Wawa Subprovince and those within the Chapleau block. The authors also recognize the extremely high metamorphic grades within the Chapleau block, but consider the grades to be not all that much higher than those recorded at the Wawa–Quetico subprovince boundary to the west (Percival 1989). A complicating factor with a subprovince boundary model for the Chapleau block is the need for both a sinistral and dextral fault to explain the exotic nature of the block. Whereas there is map pattern and structural evidence of sinistral motion for the system of faults on the east side of the block, evidence of a complementary system of dextral strike-slip faults at the western side of the Kapuskasing Structural Zone is less adequately constrained (e.g., the Wakusimi and Lepage faults; Thurston et al. 1977; Leclair 1990).

WAWA–ABITIBI SUBPROVINCE RELATIONS

For some time these 2 subprovinces were considered as one, cut by the Kapuskasing Structural Zone (Percival and Card 1985; Card and Ciesielski 1986). Jackson and Sutcliffe (1990) showed that these 2 subprovinces had similar aspects of their tectonic history, and other aspects that were in contrast. Both subprovinces contain greenstone belts, but only the Abitibi Subprovince contains komatiitic volcanic rocks as young as 2.7 Ga. Much older granitoid and volcanic rocks (approximately 2.89 Ga) occur within the Wawa Subprovince but have not been recognized in the Abitibi Subprovince. In the southern Abitibi and Batchawana greenstone belts, not until the final volcanic outburst was there evidence of an older, sialic substrate which furnished xenocrystic zircon. In the Wawa Subprovince, there is consistent but patchy evidence of an older substrate, as unconformable relationships, or REE and isotopic patterns that demonstrate the melting of older and evolved crust.

If the 2 subprovinces, or their components, did to some degree evolve independently, then the nature of the Kapuskasing Structural Zone may be more than an intracontinental thrust (Percival 1990). Alternatively, the zone may be a suture between 2 subprovinces that coalesced late in their evolution. This weak zone may subsequently have been reactivated as an oblique thrust.

OUTSTANDING PROBLEMS

A number of problems have been described and illustrated in this chapter, but the following are highlighted here as examples of aspects of the geology where ignorance is still dominant over data:

1. Additional precise age determinations are required to substantiate the recognition and correlation of assemblages.
2. The granitoid terranes between greenstone belts are relatively unknown, yet may be presumed to have had significant control on the processes of greenstone belt evolution.
3. The timing and distribution of folds, planar and linear fabrics, and shear zones, their relative and absolute chronology, is only imperfectly known in all but the Shebandowan greenstone belt, and there, only for a relatively small region.
4. What were the original relations between the Abitibi and Wawa subprovinces?

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Southern Province and Proterozoic Events

Chapter 13

Proterozoic Geology of Ontario: Introduction

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The following 5 chapters describe the Proterozoic rocks of the Southern Province of Ontario. The tectonic framework for the Proterozoic geology of North America has been developed in reviews by Lewry et al. (1987) and Hoffman (1988, 1989). The Proterozoic rocks described here occur as part of an anastomosing network of orogenic belts welding disparate Archean cratons (*see* Archean Geology of Ontario: Introduction); this network is termed the Pan American System (Lewry et al. 1987) and is represented in northern Ontario by the Trans–Hudson Orogen and in central Ontario by the Penokean Orogen (Figure 13.1). The Southern Province is composed of rocks of the Penokean Orogen. In Ontario, the Trans–Hudson Orogen is cloaked by Phanerozoic rocks, except for the Sutton Inlier; the Sutton Inlier is described in the tectonic summary chapter (Williams, Stott, Thurston et al, this volume).

At the southern margin of the Superior Province, the Southern Province consists of Paleoproterozoic, 2.4 to 2.2 billion-year-old siliceous, continental margin sedimentary rocks of the Huronian Supergroup (Cobalt Embayment and the Penokean Fold Belt). South of Lake Superior, the edge of the Superior Province is overlain by the Marquette Range Supergroup, a 2.2 to 1.7 billion-year-old deformed, rifted, passive margin sequence (Mille Lacs and Chocolay groups) overlain by the Animikie Group, a synorogenic foredeep sequence (Hoffman 1987) which extends northward into Ontario. The Huronian Supergroup is affected by a poorly documented, less than 2.2 billion-year-old deformation and the 1.8 billion-year-old compressional Penokean Orogeny (Figure 13.2).

The 1.85 billion-year-old Sudbury Igneous Complex and associated Whitewater Group sedimentary rocks lie along the northern margin of the Penokean Fold Belt, east of Lake Huron (*see* Figure 13.2). The origin of the Sudbury Igneous Complex and the associated Whitewater Group breccias and sedimentary sequence is controversial; some support a meteorite impact origin and others advocate an endogenic petrogenetic model (Pye et al. 1984).

The Mesoproterozoic is developed only in the Lake Superior region where it is characterized by anorogenic magmatism, manifested by 1.5 Ga, saturated to undersaturated plutons (e.g., English Bay pluton; *see* Sutcliffe, this volume). A minor correlative unnamed pyroclastic unit occurs in the Lake Nipigon area and overlying sedimentary rocks may be correlative with the Sibley Group, an epicratonic cover sequence.

Rifting produced an arcuate structure over 2000 km long (King and Zeitz 1971; Halls 1978), the Midcontinent Rift, which, in the Lake Superior region, is filled with up to 30 km thickness of Keweenawan Supergroup basalts and sedimentary rocks. Related intrusive rocks include layered mafic plutons, diabase sills and dike swarms, and mafic to felsic ring complexes. The rifting event is also reflected by the development of alkalic complexes emplaced along the older Kapuskasing Structural Zone, a north- to northeast-trending, west-dipping thrust, cutting the Superior Province craton over a length of 800 km.

This section of the volume deals with 2 aspects of the Proterozoic in Ontario: 1) description of the major units of the Southern Province in the Great Lakes region on the southern flank of the Archean craton; and 2) the response of the Superior craton to Proterozoic orogenesis and reactivation of the Archean structures, reflected in the development of Proterozoic dike swarms and alkalic complexes.

The variable data base and the economic importance of areas have necessitated variations in the approach to the chapters in this section of the volume.

1. The chapter on the Huronian Supergroup and Cobalt Embayment (*see* Bennett et al., this volume) describes Proterozoic rocks of the Lake Huron area. The chapter emphasizes the documentation of a revised Huronian stratigraphy and integration of the revised stratigraphy into a tectonic context. As well, metasomatic cobalt-silver mineralization of the Cobalt mining camp and paleoplacer gold and uranium mineralization near Elliot Lake are placed into a tectonic context.
2. The chapter on the Sudbury Structure (*see* Dressler et al., this volume) provides a concise overview of the Sudbury Igneous Complex rock types and their field relationships, petrography and genesis, including the latest radiogenic isotopic and geochronologic results. The description of the Sudbury Igneous Complex is background essential to a concise description of the premier Canadian nickel producer. The complex relationship between the Sudbury Igneous Complex, unusual breccias and the overlying sedimentary rocks are comprehensively described. This adds to the complexity of genetic interpretations.
3. The chapter on the Proterozoic geology of the Lake Superior area (*see* Sutcliffe, this volume) describes Proterozoic sedimentary and intrusive rocks of the Lake Superior region, providing an overview of some of the least deformed Precambrian supracrustal rocks in the province. The chapter details the igneous petrology

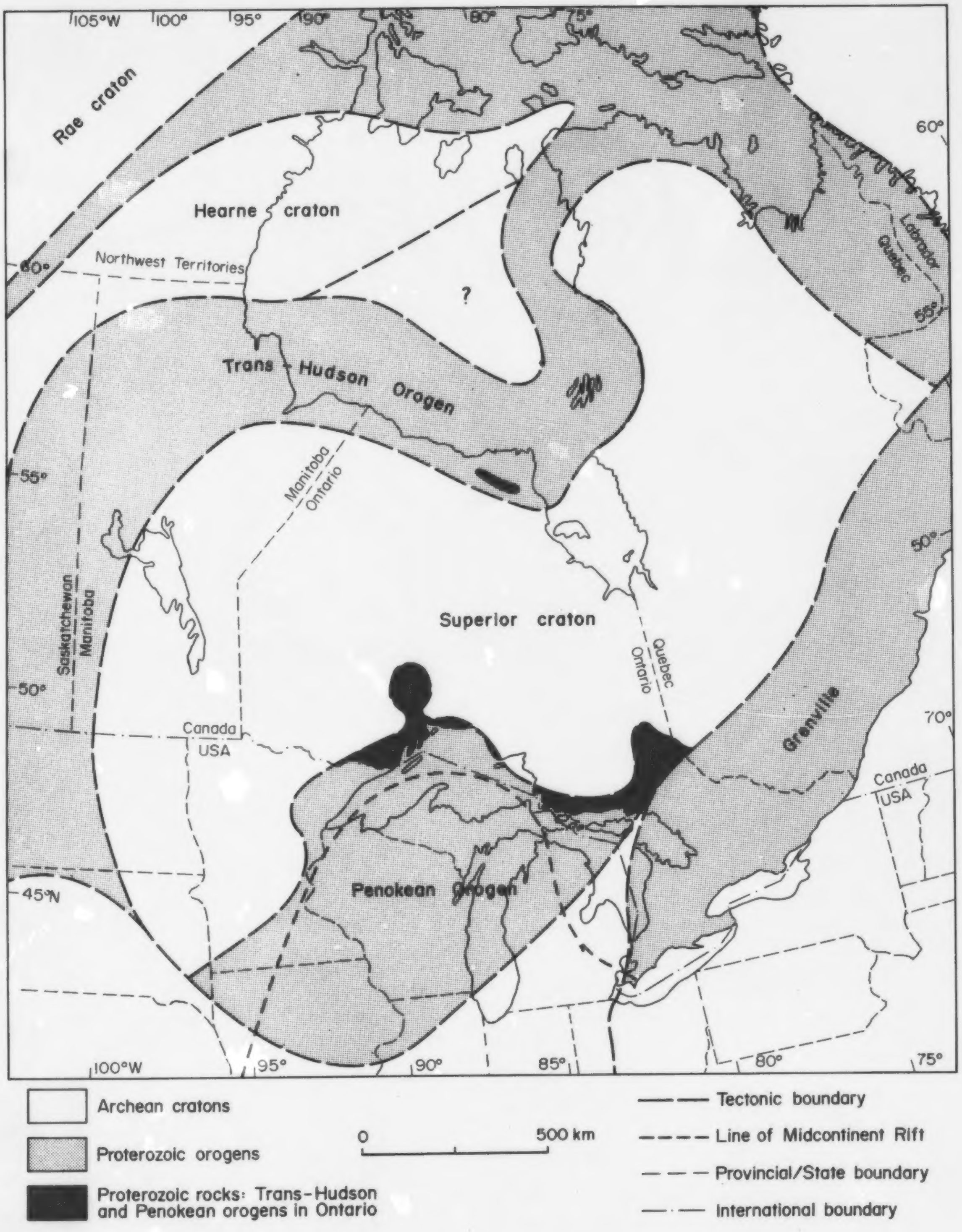
105°W
100°
95°
90°
80°
75°
Rae craton
Hearne craton
60°
Northwest Territories
?
Trans-Hudson Orogen
55°
Manitoba
Ontario
Saskatchewan
Manitoba
Superior craton
Quebec
Ontario
50°
Canada
USA
Grenville
70°
Canada
USA
Labrador
Quebec
55°
50°
Penokean Orogen
45°N
100°W
95°
90°
85°
80°
75°
Archean cratons
Proterozoic orogens
Proterozoic rocks: Trans-Hudson and Penokean orogens in Ontario
0
500 km
Tectonic boundary
Line of Midcontinent Rift
Provincial/State boundary
International boundary

Figure 13.1. Sketch map showing the Proterozoic orogens of Ontario flanked by Archean cratons (*modified from* Hoffman 1989).

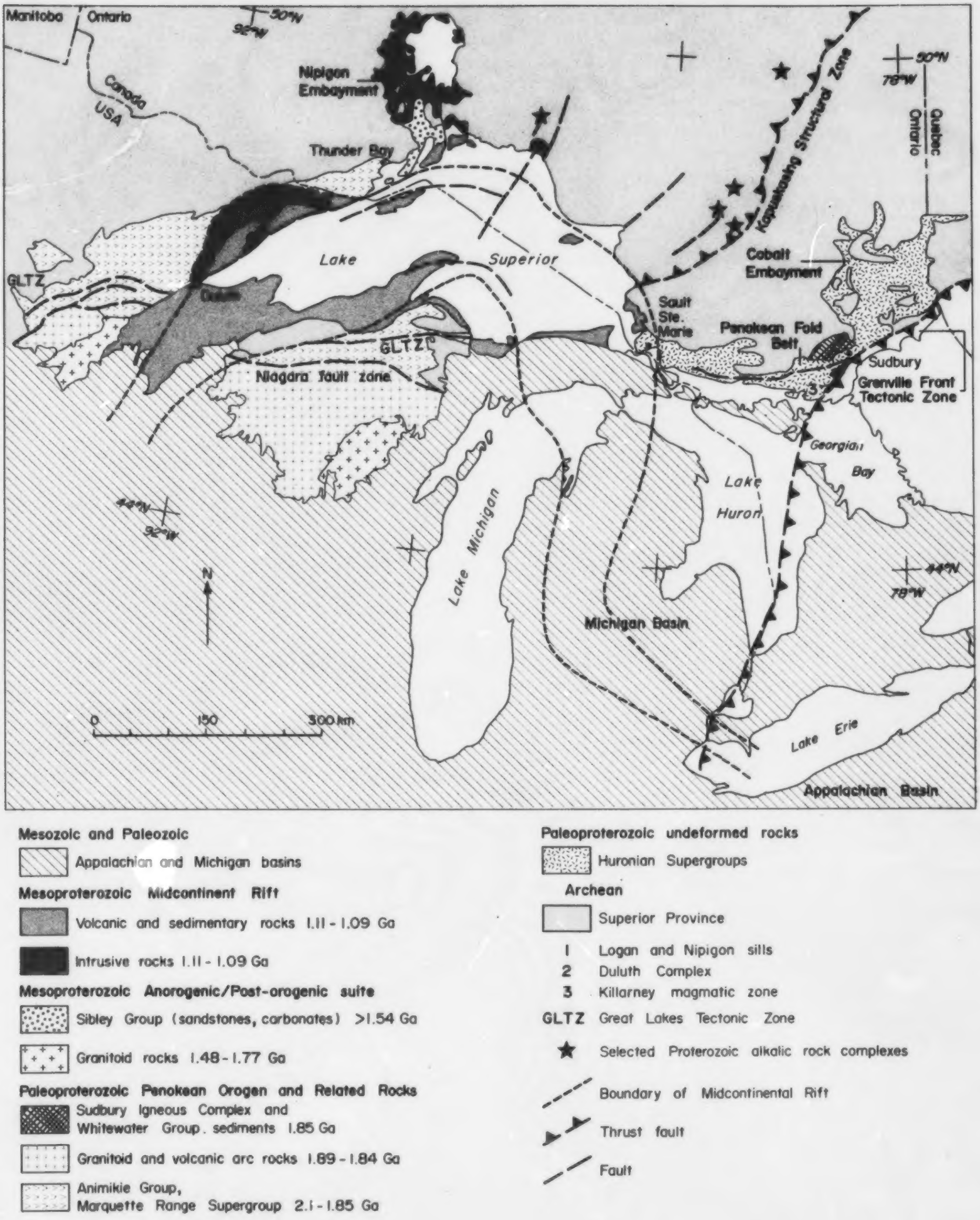
Manitoba
Ontario
50°N
92°W
Canada
USA
Nipigon Embayment
Thunder Bay
Kapuskasing Structural Zone
50°N
78°W
Quebec
Ontario
Lake
Superior
GLTZ
Duluth
Cobalt Embayment
Sault Ste. Marie
Penokean Fold Belt
Sudbury
Grenville Front Tectonic Zone
GLTZ
Niagara fault zone
Georgian Bay
Lake Huron
Lake Michigan
44°N
92°W
N
44°N
78°W
Michigan Basin
0
150
300 km
Lake Erie
Appalachian Basin
Mesozoic and Paleozoic
Appalachian and Michigan basins
Mesoproterozoic Midcontinent Rift
Volcanic and sedimentary rocks 1.11 - 1.09 Ga
Intrusive rocks 1.11 - 1.09 Ga
Mesoproterozoic Anorogenic/Post-orogenic suite
Sibley Group (sandstones, carbonates) >1.54 Ga
Granitoid rocks 1.48 - 1.77 Ga
Paleoproterozoic Penokean Orogen and Related Rocks
Sudbury Igneous Complex and Whitewater Group. sediments 1.85 Ga
Granitoid and volcanic arc rocks 1.89 - 1.84 Ga
Animikie Group, Marquette Range Supergroup 2.1 - 1.85 Ga
Paleoproterozoic undeformed rocks
Huronian Supergroups
Archean
Superior Province
1 Logan and Nipigon sills
2 Duluth Complex
3 Killarney magmatic zone
GLTZ Great Lakes Tectonic Zone
Selected Proterozoic alkalic rock complexes
Boundary of Midcontinental Rift
Thrust fault
Fault

Figure 13.2. Sketch map of the major Proterozoic units of the Great Lakes region (*modified from* Sutcliffe, this volume).

of the mafic rocks related to the Keweenawan rifting event, and integrates petrologic data with recent seismic and geochronologic data to provide an up-to-date synthesis.

4. The chapter describing Proterozoic mafic dike swarms in the Superior Province (*see* Osmani, this volume) provides an overview of the major Proterozoic diabase dike swarms cutting the Superior Province. The emphasis is on the distribution, petrographic and geochemical character of the dike swarms and their tectonic setting. The chapter includes much new data on the extent of particular dike swarms and the constraints offered by dike analysis in tectonic reconstruction.
5. The chapter on alkalic complexes (*see* Sage, this volume) discusses intrusions within the Superior Province, which range in age from Proterozoic through to Paleozoic. The chapter details the petrographic, geochemical and isotopic variations among these alkalic complexes.

This section of the volume illustrates the enormous strides made in correlation of the Precambrian of the Great Lakes region since the advent of high-precision U-Pb dating. It is now clearer than heretofore what the relationships are between the Huronian Supergroup and the Marquette Range Supergroup. The relationship between the Huronian Supergroup continental margin prism and the foredeep sequence of the Animikie Group is now clear (cf., Hoffman 1987; 1989), permitting a sophisticated tectonic interpretation (*see* Bennett et al., this volume; Barovich et al. 1989). New information on the Sudbury Structure is used to constrain a comprehensive review of these important ore deposits. The evidence still leaves the origin of the structure a question—is it related to impact or an endogenic event?

The petrologic and seismic data on the Keweenawan Midcontinent Rift indicate the enormous size of the feature (Cannon et al. 1989). The thickness of the basaltic section identifies the Keweenawan basalts as perhaps the earth's largest plateau basalt eruption. Integration of this geochronologically constrained synthesis with future transborder magnetic and gravity studies will allow tracing of major units across the border. This will permit a new generation of interpretation of Proterozoic events in the midcontinent area.

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Chapter 14

The Huronian Supergroup and Associated Intrusive Rocks

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Abstract

The eastern part of the Southern Province consists of the Paleoproterozoic Huronian Supergroup (about 2500 to 2220 million years old) which occurs in an area extending from the east shore of Lake Superior, along the north shore of Lake Huron to Sudbury, and beyond, toward Noranda in northwestern Quebec. The supergroup is subdivided into 4 stratigraphic groups. The oldest, the Elliot Lake Group, consists of volcanic rocks and clastic sedimentary rocks, and contains significant uranium deposits. The 3 other groups, from bottom to top, are the Hough Lake, Quirke Lake and Cobalt groups. Each is characterized by a sedimentary cycle in which conglomerate, of probable glacial origin, is overlain by mudstone, siltstone and coarse arenite. The Quirke Lake Group is exceptional for its calcareous sedimentary rocks.

In the 3 lower groups, reddish colours are scarce; by comparison, some rock units of the Cobalt Group exhibit a red colouration that correlates with the preservation of detrital uraninite and pyrite. Distribution, chemistry and structure of paleosols interspersed within the stratigraphy of the Huronian Supergroup indicate a change from an oxygen-poor to an oxygen-bearing atmosphere during deposition.

East-trending, north-verging synclines and anticlines and associated south-dipping thrust faults; conjugate sets of northeast- and northwest-trending, strike-slip faults; and the east-trending, dextral, strike-slip Murray Fault all indicate oblique, northward compression across the Penokean Fold Belt. Within the Penokean Fold Belt, the intensity of deformation decreases northwards; deformation in the Cobalt Embayment is restricted to open folding and normal faulting.

Supracrustal rocks were generally subjected to low-grade metamorphism, but amphibolite-facies mineral assemblages are known to exist within the Penokean Fold Belt from immediately west of Sudbury to as far west as Spragge.

Several intrusive rock suites, such as East Bull Lake Gabbro-Anorthosite Intrusion, Nipissing diabase, Croker Island and Cutler intrusions, are spatially and perhaps genetically associated with both the supracrustal development and subsequent dynamothermal metamorphism of the Huronian Supergroup.

We propose that formation of the Huronian Supergroup was initiated on a passive margin and subsequently became involved in a convergent tectonic regime.

The Huronian Supergroup is known for its uranium deposits. Silver, cobalt, copper, minor amounts of nickel and platinum group elements occur in deposits associated with diabase sills intruding the sedimentary rocks of the supergroup.

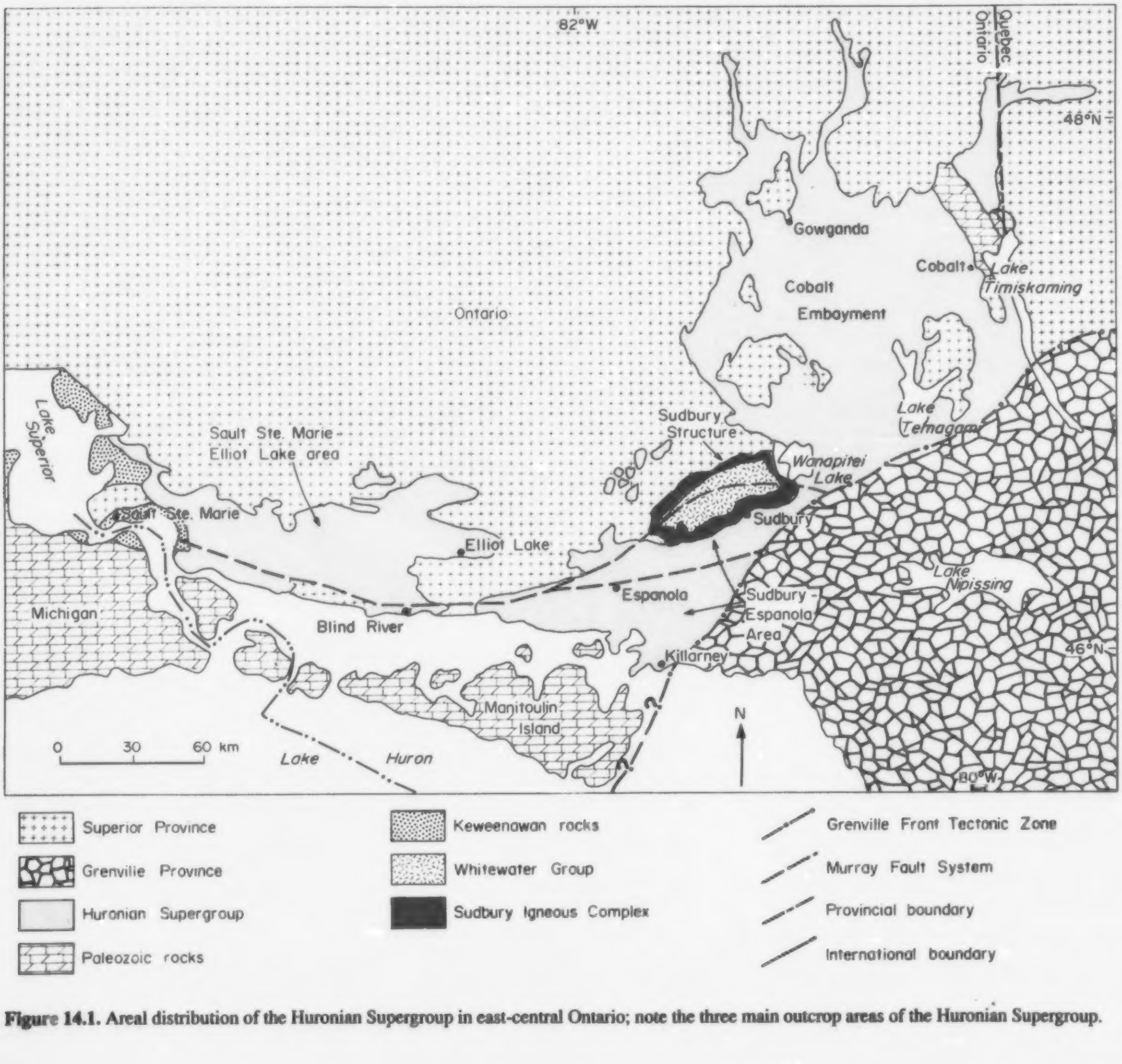
INTRODUCTION

All sedimentary and volcanic rocks that are within the Huronian Supergroup are metamorphosed. Although omitted in discussion, prefix "meta-" is nonetheless implied. The Paleoproterozoic Huronian Supergroup forms part of the Southern structural province of the Canadian Shield. The supergroup consists of sedimentary and subordinate volcanic rocks which lie unconformably upon Archean rocks of the Superior Province and form a belt, the Penokean Fold Belt, that is up to 160 km wide. Rocks in the supergroup extend from the east shore of Lake Superior, eastward along the north shore of Lake Huron to Sudbury, and northeastward, across the Cobalt Embayment, to the Noranda area in northwestern Quebec, a distance of about 450 km (Figure 14.1). Southwest of Sudbury, the Huronian Supergroup attains its greatest thickness of 12 km. The sequence thins northward (Figure 14.2) due to the wedging out of basal units, a thinning of clastic units, and erosion within the sequence (Frarey and Roscoe 1970). Stratigraphic thickness estimates for areas north of the Murray Fault System, in areas of low to medium metamorphic grade, are well constrained; south of the Murray Fault System, tectonic thickening renders estimates less precise.

The Huronian Supergroup (Figure 14.3) consists of 4 groups, which in ascending stratigraphic order are: the Elliot Lake, Hough Lake, Quirke Lake and Cobalt groups. This stratigraphic order has been formalized by Robertson et al. (1969a) who also defined the type locations for the various formations of the supergroup. A spectrum of opinion exists on the tectonic significance of the subdivision of the Huronian Supergroup into its constituent 4 groups. Numerous previous investigations have classified the group-level contacts within the Huronian Supergroup as conformable contacts (e.g., references by Card). However, Zolnai et al. (1984) categorized the Supergroup as 4 unconformity-bounded sequences. The Elliot Lake Group at the base of the Huronian Supergroup contains Witwatersrand-type uranium deposits. It is the only group that contains volcanic rocks and turbidites and has significant lateral variations in thickness and rock type. It does not display the threefold paleoenvironmental subdivision characteristic of the overlying Hough Lake, Quirke Lake and Cobalt groups. These latter 3 groups form 3 sedimentary cycles of conglomerate, overlain by either mudstone, siltstone or carbonate, and are capped by coarse, cross-bedded sandstone (Roscoe 1969). Many paleocurrent studies have shown that the source area for sedimentary rocks of the Huronian Supergroup lay north to west of the area of deposition, with a source area to the northwest having predominated. To the northeast of Sudbury, paleocurrent directions are more diverse, but are most commonly from the north (Long 1976; Rice 1987, 1988; Debicki 1990).

Several igneous intrusions are spatially associated with rocks of the Huronian Supergroup: 1) the East Bull Lake Gabbro-Anorthosite Intrusion, 2) granitic plutons of the Sudbury area, 3) the Nipissing intrusive rocks (2219 Ma, *see* Bedrock Geology of Ontario, maps 2541 to 2545, map case), 4) the Croker Island and Manitoulin Island complexes, 5) the Cutler monzonite-tonalite, and 6) the olivine diabase of the Sudbury dike swarm (1238 Ma, *see* Bedrock Geology of Ontario, maps 2541 to 2545, map case). Not all of these intrusions are described in this review.

The form of the depositional basin for the Huronian Supergroup was influenced by Archean tectonic processes and a Paleoproterozoic extension event. Zolnai et al. (1984) have suggested that rocks of the Elliot Lake Group, in the southern part of the region underlain by the Huronian Supergroup, were deposited within east-trending grabens separated by areas where sediment was not deposited, or has been removed, but detailed corroboration of that model is not complete. Rocks of the Cobalt Group, in the northern part of the areal extent of the Huronian Supergroup, in the vicinity of Kirkland Lake, were partly deposited within a north-trending graben of Archean age (Powell et al. 1990)



82°W
Ontario
Quebec
48°N
Gowganda
Cobalt
Lake Timiskaming
Cobalt Embayment
Ontario
Lake Temagami
Sudbury Structure
Wanapitei Lake
Sudbury
Lake Superior
Sault Ste. Marie - Elliot Lake area
Sault Ste. Marie
Elliot Lake
Espanola
Lake Nipissing
Sudbury - Espanola Area
Michigan
Blind River
Killarney
46°N
Manitoulin Island
N
0 30 60 km
Lake Huron
80°W
Superior Province
Grenville Province
Huronian Supergroup
Paleozoic rocks
Keweenawan rocks
Whitewater Group
Sudbury Igneous Complex
Grenville Front Tectonic Zone
Murray Fault System
Provincial boundary
International boundary

Figure 14.1. Areal distribution of the Huronian Supergroup in east-central Ontario; note the three main outcrop areas of the Huronian Supergroup.

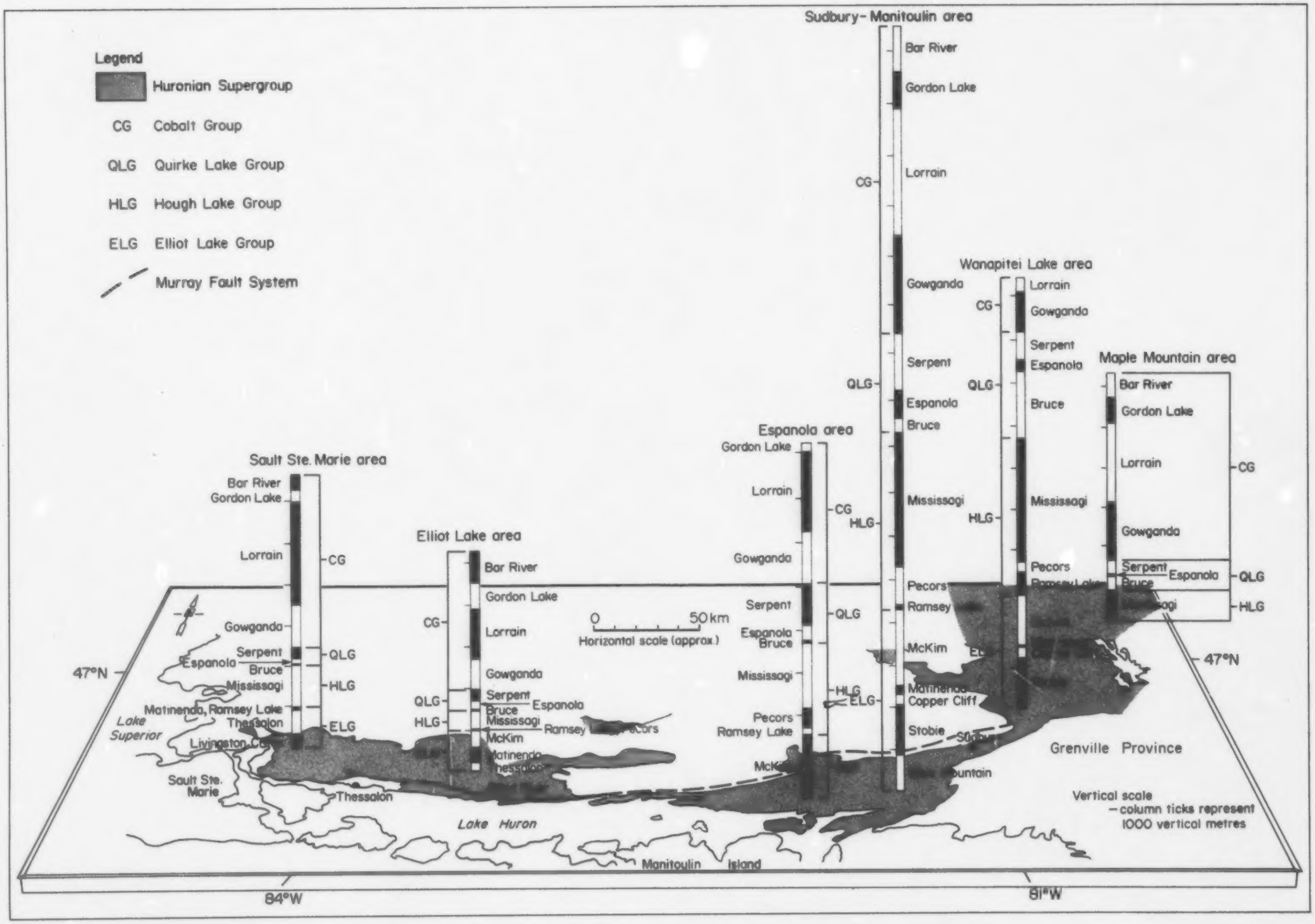
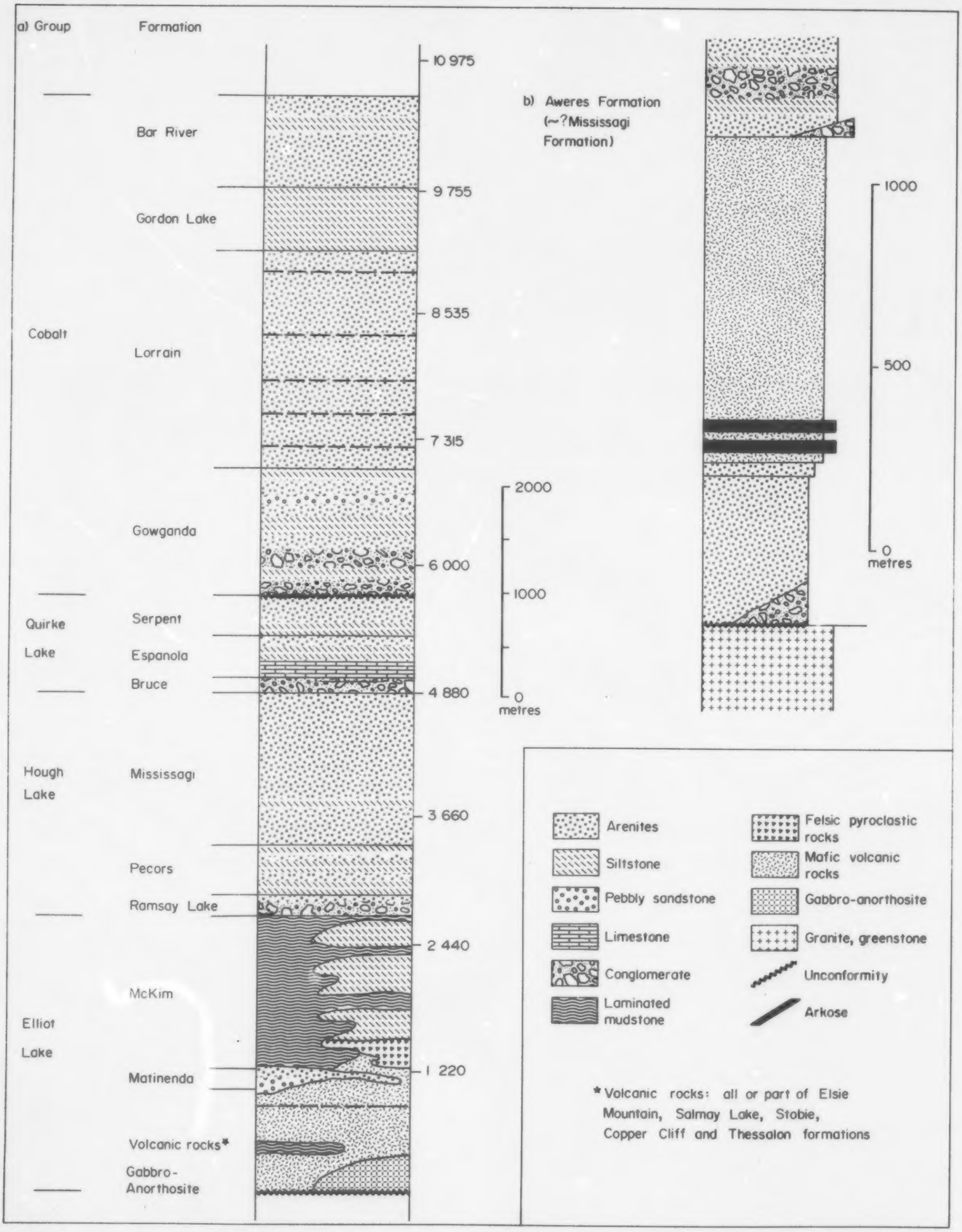

Legend
Huronian Supergroup
CG Cobalt Group
QLG Quirke Lake Group
HLG Hough Lake Group
ELG Elliot Lake Group
Murray Fault System
Sault Ste. Marie area
Bar River
Gordon Lake
Lorrain
Gowganda
Serpent
Espanola
Bruce
Mississagi
Matinenda, Ramsey Lake
Thessalon
Livingston Cr.
CG
QLG
HLG
ELG
Elliot Lake area
Bar River
Gordon Lake
Lorrain
Gowganda
Serpent
Espanola
Bruce
Mississagi
Ramsey
Pecors
McKim
Matinenda
Thessalon
Espanola area
Gordon Lake
Lorrain
Gowganda
Serpent
Espanola
Bruce
Mississagi
Pecors
Ramsey Lake
McKim
Sudbury- Manitoulin area
Bar River
Gordon Lake
Lorrain
Gowganda
Serpent
Espanola
Bruce
Mississagi
Pecors
Ramsey
McKim
Matinenda
Copper Cliff
Stobie
Sudbury
Mountain
Wanapitei Lake area
Lorrain
Gowganda
Serpent
Espanola
Bruce
Mississagi
Pecors
Ramsey Lake
Maple Mountain area
Bar River
Gordon Lake
Lorrain
Gowganda
Serpent
Espanola
Bruce
Mississagi
0
50 km
Horizontal scale (approx.)
47°N
47°N
84°W
81°W
Lake Superior
Sault Ste. Marie
Thessalon
Lake Huron
Manitoulin Island
Grenville Province
Vertical scale
– column ticks represent
1000 vertical metres

Figure 14.2. Block diagram illustrating the variation in stratigraphic thickness of units in the Huronian Supergroup, Southern Province.


a) Group
Formation
10 975
Bar River
9 755
Gordon Lake
8 535
Cobalt
Lorrain
7 315
Gowganda
6 000
Quirke Lake
Serpent
Espanola
Bruce
4 880
Hough Lake
Mississagi
3 660
Pecors
Ramsay Lake
2 440
McKim
Elliot Lake
Matinenda
1 220
Volcanic rocks*
Gabbro-Anorthosite
2000
1000
0 metres
b) Aweres Formation (~?Mississagi Formation)
1000
500
0 metres
Arenites
Siltstone
Pebbly sandstone
Limestone
Conglomerate
Laminated mudstone
Felsic pyroclastic rocks
Mafic volcanic rocks
Gabbro-anorthosite
Granite, greenstone
Unconformity
Arkose

*Volcanic rocks: all or part of Elsie Mountain, Salmay Lake, Stobie, Copper Cliff and Thessalon formations

Figure 14.3. Generalized stratigraphic section of a) the Huronian Supergroup, and b) the Aweres Formation. The Aweres Formation in the Sault Ste. Marie area is thought to be equivalent to parts of the Mississagi Formation.

based, in part, on facies variation within Cobalt Group units (Rice 1988; Debicki 1990).

The rocks of the Huronian Supergroup are dated at between 2219 Ma and slightly more than 2450 Ma. The younger age is that of Nipissing gabbro, diabase and granophyre that intruded the Huronian Supergroup (Corfu and Andrews 1986); the older age date was determined from the rhyolitic Copper Cliff Formation (Krogh et al. 1984), a rock unit of the Elliot Lake Group occurring near the base of the Huronian Supergroup. The 2480 million-year-old East Bull Lake Gabbro-Anorthosite Intrusion (Krogh et al. 1984) is generally believed to be related to volcanism in the Huronian Supergroup, and provides a constraint for the age of the earliest rocks in the Huronian Supergroup.

Rocks of the Huronian Supergroup have been affected by an orogeny prior to intrusion of the Nipissing diabase (Zolnai et al. 1984) at about 2.2 Ga (Corfu and Andrews 1986), and by the Penokean Orogeny. The age of the Penokean Orogeny has been variously estimated as being:

1. 1.7 to 1.9 Ga (Cannon 1973)
2. 1.7 to 1.75 Ga (Hurst and Farhat 1978)
3. 1.89 to 1.83 Ga (Bickford et al. 1986)
4. 1.8 to 1.9 Ga (Medaris 1983)
5. 1.86 to 1.835 Ga (Hoffman 1989)

No clear choice between the age ranges listed above is possible, based upon Ontario data. During the Penokean event, the rocks of the Huronian Supergroup were subjected to low-pressure, dynamothermal metamorphism.

Paleosols have been observed at several locations in areas underlain by rocks of the Huronian Supergroup and provide some constraints on the type of weathering and atmospheric conditions which existed in the Paleoproterozoic. Fluvial and shallow-marine rocks do not exhibit a red colour due to the anaerobic environment of deposition. Most researchers have interpreted the lack of red colour in rocks of the lower part of the Huronian Supergroup as being due to an oxygen-deficient atmosphere which resulted in a reduction of Fe_2O_3 to FeO and the preservation of uraninite and pyrite during deposition of the sedimentary rocks of the lower Huronian Supergroup (Roscoe 1969, 1973; Frarey and Roscoe 1970). However, recent work (Maynard et al. 1991) suggests that intense chemical weathering may have been an alternative mechanism for the preservation of detrital uraninite and pyrite.

The structural geology of the Huronian strata between Sault Ste. Marie and Sudbury (Penokean Fold Belt) is characterized by large, regional, east-trending folds and faults. In the Cobalt Embayment, the strata of the Cobalt Group have been little affected by deformation other than by normal faults and large open folds. The earliest deformation affecting the Huronian Supergroup is marked by open folding which preceded the deposition of the Gowganda Formation (Card et al. 1973; Dressler 1977) and the intrusion of the Nipissing diabase at 2.2 Ga. The deformational events of the Penokean Orogeny produced most major fold and fault structures of the Huronian Supergroup.

The rocks of the Matinenda Formation, part of the lower Huronian Supergroup, are well known for their uranium deposits which are found in pyritic quartz-pebble conglomerate interbedded with arkosic arenite.

This chapter provides a summary of what is presently known about the supergroup. It draws heavily on the work of the most recent investigators, but does not review all of the literature available.

EARLY GEOLOGICAL INVESTIGATIONS

Rocks of the Huronian Supergroup were amongst the first to be described from the Canadian Shield; they underlie a relatively easily accessible part of Ontario, along travel routes used by early European traders and settlers. The first description of what is now known as the Huronian Supergroup dates back to the early 1820s when a physician, J.J. Bigsby (1821, 1863), reported on sedimentary rocks he had observed along the north shore of Lake Huron.

The discovery of copper at Bruce Mines, in 1843, led to geological investigations and mapping between 1847 and 1858, mainly in the Blind River area, by Alexander Murray and W.E. Logan of the Geological Survey of Canada. Logan (1863) referred to the sequence of sedimentary rocks, which he recognized as unconformably overlying granitic rocks, as the "Huronian Series". Logan's recognition of this unconformity represents the first stratigraphic subdivision of Precambrian rocks anywhere in the Canadian Shield (Roscoe 1969).

Intensive and systematic investigations of the Huronian sedimentary sequences began in the early years of the twentieth century. Coleman (1905a, 1905b), Collins (1914a, 1917, 1925), Quirke (1917, 1921, 1926) and Quirke and Collins (1930) were the pioneers of research dedicated to the "Huronian Series". Collins (1925), in a comprehensive report and map of the Blind River area, put forward a lithostratigraphic subdivision for the Huronian sequences that, in large part, is similar to the one accepted today. Later, Collins (1936) worked on the Huronian sequences in the Sudbury area where, earlier, Coleman (1905a, 1905b, 1914),in the course of mapping the Sudbury Igneous Complex, had provided a stratigraphic subdivision of the adjacent Huronian rocks. Collins (1914a) had correlated the rocks south of the Sudbury Igneous Complex with the Archean "Timiskaming Series" (*see* Jackson and Fyon, this volume), a correlation not currently accepted.

Remapping of the Huronian sequence had commenced with Thomson (1952) and Abraham (1953), but it was the discovery of uranium deposits in the Elliot Lake area during the 1950s that renewed a general interest in the Huronian Supergroup. Geological mapping and research by government survey parties and intensive investigations by university staff resulted in the generation of a voluminous literature (*see* the annotated bibliography on the Huronian Supergroup by Junnila 1987).

ELLIOT LAKE GROUP

The Elliot Lake Group is lowermost in the Huronian Supergroup and contains uranium deposits. For descriptive purposes, the group is subdivided in this chapter into 3 topics for discussion: 1) sedimentary units of the lower Elliot Lake Group, 2) volcanic rocks of the Elliot Lake Group, and 3) sedimentary units of the upper Elliot Lake Group. Stratigraphic thicknesses are shown in Figure 14.4.

The Elliot Lake Group differs from the 3 other Huronian groups in that it contains volcanic rocks and significant uranium deposits, and does not have formations which contain a conglomerate-mudstone-sandstone sequence of deposition. The distribution of the volcanic rocks of the Huronian Supergroup is shown in Figure 14.5.

A sedimentary sequence, which underlies the volcanic rocks in the west (Livingstone Creek Formation), is absent from the eastern parts of the Huronian Supergroup's areal extent.

The lower contact of the Elliot Lake Group is an unconformity marked by the widespread development of paleosols (summarized by Bennett et al. 1990). The upper contact of the group has been variously interpreted as conformable (Frarey 1977) or possibly unconformable (Card and Palonen 1976). A disconformity, marked by a paleosol, is recognized in the western part of the area where rocks of the Huronian Supergroup occur, but it is not recognized in the eastern part. Correlations between map units of the Elliot Lake Group at the extreme western and eastern ends of the outcrop area of the Huronian Supergroup are tentative.

Sedimentary Units of the Lower Elliot Lake Group

Sedimentary rocks at the base of the Huronian Supergroup are assigned to the Livingstone Creek Formation. They are significant for they allow an interpretation of the earliest paleotectonic and depositional environments in the Huronian sequence (*see* A Tectonic Model for the Huronian Supergroup).

LIVINGSTONE CREEK FORMATION

The Livingstone Creek Formation—the Driving Creek formation of McConnell (1927)—is known only from the western and central parts of the area underlain by the Huronian Supergroup. Near Sault Ste. Marie, it has a maximum thickness of 400 m, whereas northeast of Thessalon the formation is 300 m thick, as has been indicated by drilling. By contrast, outcrops of the Livingstone Creek Formation indicate a thickness of only 100 m (Frarey 1977). At Crazy Lake in Nicholas Township, in the central parts of the outcrop area of the Huronian Supergroup, the formation is only 150 m thick. In the area around Morin and Otter townships, subsurface units correlative with the Livingstone Formation have been observed in drill core.

The Livingstone Creek Formation consists almost entirely of fine- to medium-grained arenites to wackes, grey arkose, subarkose, minor wacke and subwacke. Also present is clast-supported, polymictic conglomerate (Figure 14.6). Trough cross-bedding, locally with calcite-rich foresets, is commonly developed. Planar cross-bedding and normal bedding are rare.

Near Sault Ste. Marie, up to 10 m of the uppermost part of the Livingstone Creek Formation are composed of a fine- to medium-grained, well sorted, massive, quartz arenite (Bennett 1977). The lower part of the quartz arenite contains much iron-bearing carbonate (possibly ferroan dolomite) along foreset beds and along the contact with the underlying arkose.

Clast-supported, polymictic conglomerate is prominent in the lower parts of the Livingstone Creek Formation. Boulders and cobbles of grey granitic rocks and subordinate amounts of mafic metavolcanic, plutonic igneous and metamorphic rock are set in a matrix of coarse arkose or arkosic grit. Units of cross-bedded arkose are locally interbedded with the conglomerate (Frarey 1977; Bennett 1981). The compositions of the conglomerate clasts usually reflect that of the underlying rocks.

Clast-size distribution and local source rocks for the conglomerate are consistent with an interpretation that part of the formation consists of alluvial-fan deposits. However, the trough cross-bedded units are much finer grained than the arkosic matrix of the conglomerate. These generally fine-grained sandstones have a source different from that of the conglomerates and likely represent fluvial deposits.

Relative Age of the Livingstone Creek Formation

The relationship between the Livingstone Creek Formation and the overlying Matinenda Formation has been defined by field observations in the Dunns Valley area (Bennett et al. 1990). Near the southern boundary of Otter Township, sandstone of the Livingstone Creek Formation is cut by a mafic dike; the Livingstone Creek Formation is in turn overlain by arkose and quartz-pebble conglomerate of the Matinenda Formation. Beneath the Matinenda Formation, a yellowish, sericitic paleosol mantles both the Livingstone Creek sandstone and the dike. Drill holes a few kilometres southwest of this occurrence intersected Thessalon Formation basaltic flows overlying grey sandstone correlated with the Livingstone Creek Formation. The only known igneous event that is post–Livingstone Creek Formation and pre–Matinenda Formation in the Huronian Supergroup is represented by volcanic rocks in the Thessalon Formation. From these observed relationships, the dike can be temporally correlated with the Thessalon Formation. These relationships indicate that a period of volcanic activity, followed by a significant period of erosion, separated the deposition of the Livingstone Creek from that of the Matinenda Formation.

Volcanic Rocks of the Elliot Lake Group

Volcanic rocks are present at several locations within the Elliot Lake Group. Lithostratigraphic correlation between

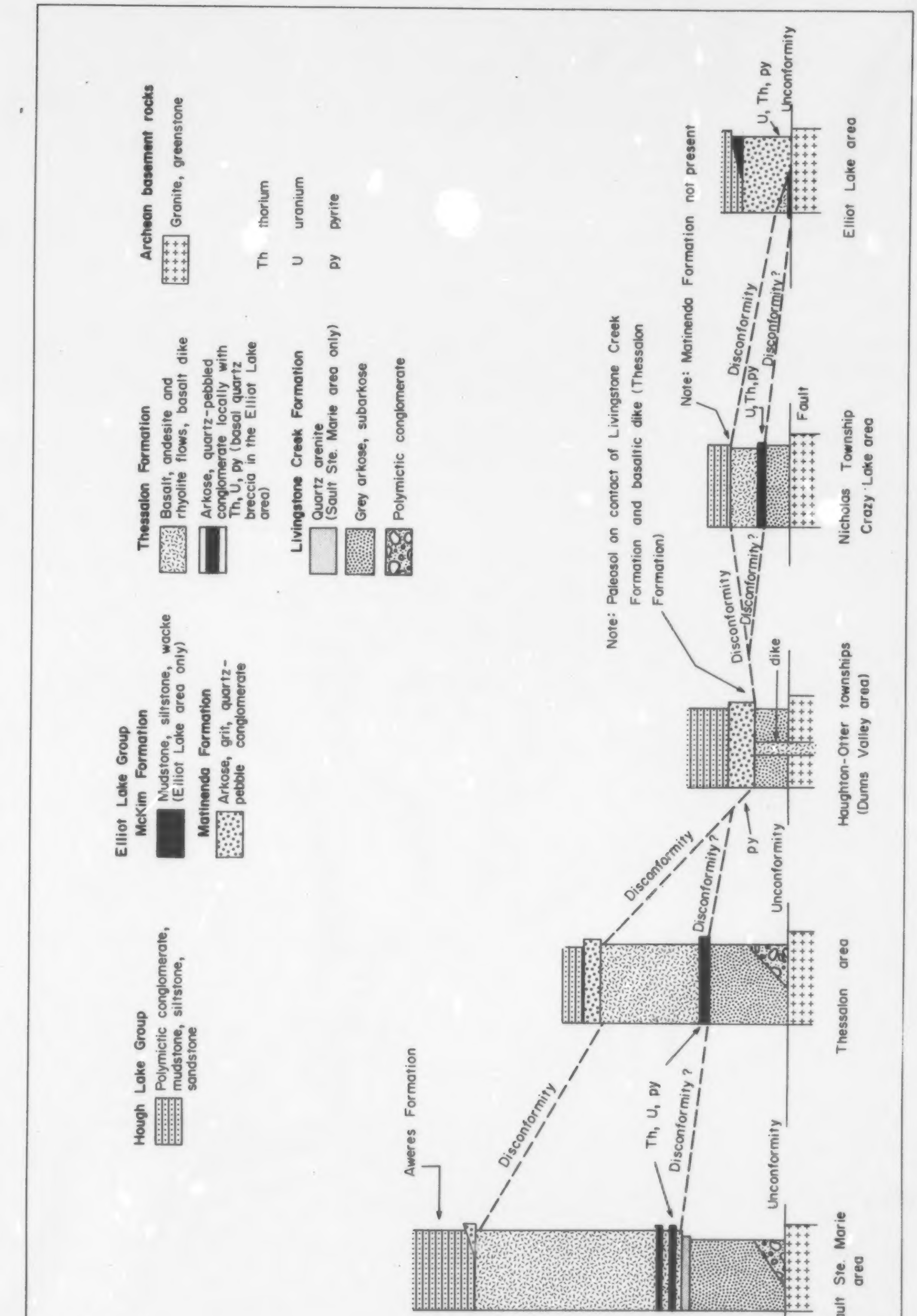

Hough Lake Group
Polymictic conglomerate, mudstone, siltstone, sandstone
Elliot Lake Group
McKim Formation
Mudstone, siltstone, wacke (Elliot Lake area only)
Matinenda Formation
Arkose, grit, quartz-pebble conglomerate
Thessalon Formation
Basalt, andesite and rhyolite flows, basalt dike
Arkose, quartz-pebbled conglomerate locally with Th, U, py (basal quartz breccia in the Elliot Lake area)
Livingstone Creek Formation
Quartz arenite (Sault Ste. Marie area only)
Grey arkose, subarkose
Polymictic conglomerate
Archean basement rocks
Granite, greenstone
Th thorium
U uranium
py pyrite
metres
1000
0
Aweres Formation
Disconformity
Th, U, py
Disconformity ?
Unconformity
Sault Ste. Marie area
Thessalon area
py
Haughton-Otter townships (Dunns Valley area)
dike
Note: Paleosol on contact of Livingstone Creek Formation and basaltic dike (Thessalon Formation)
Nicholas Township Crazy Lake area
U, Th, py
Fault
Note: Matinenda Formation not present
Elliot Lake area
U, Th, py

Figure 14.4. Stratigraphic columns and correlation for the Elliot Lake Group between the Sault Ste. Marie and Elliot Lake areas, Penokean Fold Belt.

volcanic sequences in the western part of the Penokean Fold Belt and those in the eastern part of the area, near Sudbury, is uncertain; moreover, it has not been attempted because the volcanic rocks form geographically isolated units that we now describe as the western, central and eastern parts of the Penokean Fold Belt (the Western, Central and Eastern volcanic rocks).

WESTERN VOLCANIC ROCKS OF THE PENOKEAN FOLD BELT

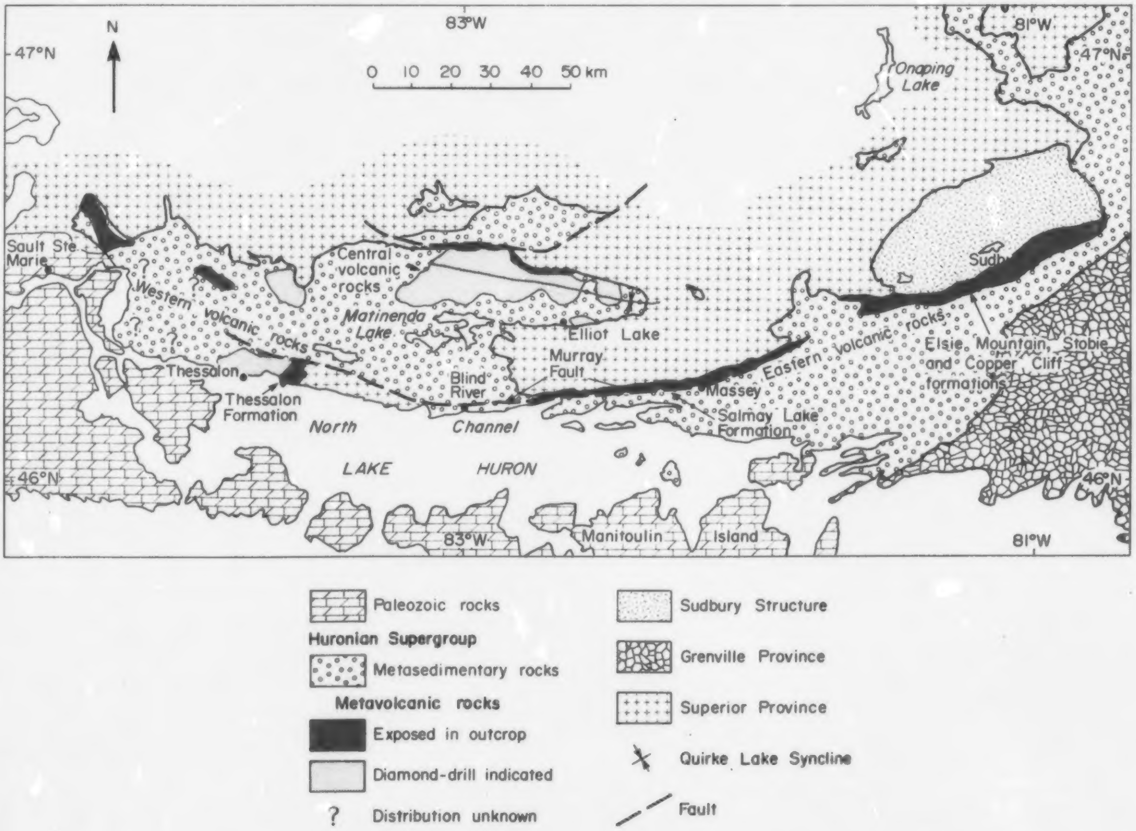
The Thessalon Formation

Mafic flows overlying the Livingstone Creek Formation in the Sault Ste. Marie and Thessalon areas, and in Aberdeen Township are named the Thessalon Formation (Frarey 1977). These volcanic rocks had been recognized for some time in the Thessalon area (Knight 1915) and were previously named the "Duncan volcanics" by McConnell (1927) near Sault Ste. Marie.

The thickness of the Thessalon Formation in the Sault Ste. Marie area is between 650 and 820 m (Frarey 1977); in central Thessalon Township it is approximately 330 m thick, and in Aberdeen Township it is approximately 1200 m thick. Drilling, south of Thessalon, into bedrock beneath Lake Huron intersected a minimum thickness of 670 m of volcanic rocks.

The Thessalon Formation is predominantly a fine- to medium-grained, amygdaloidal basalt and andesite to rhyodacite. Subordinate rhyolite flows are present near Thessalon and in Aberdeen Township. Felsic to intermediate tuffs and volcanic breccias make up a minor proportion of the formation.

The metabasalts are almost uniformly fine to medium grained, and tholeiitic. The formation displays subgreenschist- to greenschist-facies mineral assemblages. Primary clinopyroxene is prominent in some thin sections of spilitic basalt from the Sault Ste. Marie area. Quartz is a minor component of basaltic and andesitic rock types. Amygdules filled with epidote, chlorite, calcite, quartz and stilpnomelane, in complex zonal arrangements, are common in the volcanic rocks. Elongate, chlorite-filled amygdules, 1 cm or less across, are a distinctive feature of most Huronian mafic flows between the Quirke Lake Syncline and Sault Ste. Marie (Figure 14.7). Pillow structures are rare, but were observed in Duncan Township near Sault Ste. Marie. Flows with an upper crust of scoria and cross-cutting breccia are commonly filled with a fine-grained mixture of quartz and grey to red secondary albite. Welded textures are visible in some rhyolitic tuffs near Thessalon.



83°W
N
47°N
0 10 20 30 40 50 km
81°W
47°N
Onaping Lake
Sault Ste. Marie
Central volcanic rocks
Western volcanic rocks
Matinenda Lake
Elliot Lake
Murray Fault
Sudbury
Eastern volcanic rocks
Elsie Mountain, Stobie and Copper Cliff formations
Thessalon
Thessalon Formation
Blind River
Massey
Salmay Lake Formation
North Channel
LAKE HURON
46°N
83°W
Manitoulin Island
46°N
81°W
Paleozoic rocks
Huronian Supergroup
Metasedimentary rocks
Metavolcanic rocks
Exposed in outcrop
Diamond-drill indicated
? Distribution unknown
Sudbury Structure
Grenville Province
Superior Province
Quirke Lake Syncline
Fault

Figure 14.5. Distribution of metavolcanic rocks of the Huronian Supergroup, eastern Southern Province.

Figure 14.6. Clast-supported, polymictic conglomerate and grit of the Livingstone Creek Formation near the shore of Lake Huron, Thessalon Township. Scale on the photo is approximately 8 cm in length.

CENTIMETRES

Figure 14.7. Amygdaloidal metabasalt of the Thessalon Formation, Highway 129, near Thessalon.

Discontinuous units of coarse arkose and associated quartz-pebble conglomerate (pyritic and radioactive in part) are locally intercalated with flows near the base of the Thessalon Formation or are found directly below the basal flow (Hay 1963; Bennett et al. 1976; Meyer 1983).

The Thessalon Formation in the Sault Ste. Marie and Aberdeen Township areas can be subdivided, in part, on the basis of "immobile" Ni, Cr, P_2O_5 and Ti contents (Figure 14.8) into an upper, predominantly tholeiitic, basalt and a lower complex also referred to as the "diverse member". In the Sault Ste. Marie area, the lower diverse member is mainly composed of basaltic andesite, but contains rhyolite, tholeiitic basalt and high-magnesium basalt in the Aberdeen Township and Thessalon areas. Figure 14.9 shows a cation plot of metavolcanic rocks of the Thessalon Formation, illustrating the essentially tholeiitic nature of the rocks and general compositional variation. High Na:K ratios in volcanic rocks of the Thessalon Formation in the Sault Ste. Marie area indicate spilitization of the basalts (Figure 14.10).

Petrochemical studies of the Thessalon Formation in the Thessalon area (Jolly 1987) revealed that the mantle source of these volcanic rocks was metasomatically modified, possibly by Archean subduction. Jolly subdivided the volcanic rocks in the Thessalon area into an upper basaltic unit and an earlier central unit of fractionated andesite and subordinate basalt.

The overall thickness of the upper part of the Thessalon Formation (at least 700 m where best preserved in the Sault Ste. Marie area), along with its uniform basaltic character, its relative uniformity of immobile element content (Bennett, unpublished data), and the thick and generally massive character of the flows, suggest that the upper part of the Thessalon Formation is part of a flood basalt sequence. The lower diverse member probably represents eruption from central vents as evinced by the diversity of rock types (including felsic pyroclastic rocks) of relatively limited lateral extent and rapid thickness changes (Bennett 1978). Metavolcanic rocks in the Quirke Lake Syncline are termed the "Dollyberry volcanics" and consist of basaltic rocks. They are correlated by Roscoe (1969) with the volcanic rocks of the Thessalon Formation.

CENTRAL VOLCANIC ROCKS OF THE PENOKEAN FOLD BELT

Huronian volcanic rocks of the Quirke Lake Syncline in the central Penokean Fold Belt rest either on Archean rocks (Bottrill 1971; Robertson 1971) or, as observed at Crazy Lake in the northwestern part of the syncline, on Huronian sedimentary rocks. At Crazy Lake, arkose with calcite laminae is overlain by uraniferous, pyritic, quartz-pebble conglomerate, which is in turn overlain by mafic flows. The rock types and stratigraphic sequence presented here are identical to those of the Livingstone Creek and Thessalon formations of the Thessalon–Sault Ste. Marie area. The Huronian volcanic rocks of the Quirke Lake Syncline are thus correlated with the Thessalon Formation (Bottrill 1971; Bennett 1978; Robertson 1971, 1986; Bennett et al. 1990).

The Huronian volcanic rocks which outcrop along the north limb of the Quirke Lake Syncline are predominantly high-magnesium tholeiitic basalts, alkalic basalts and some massive rhyolite and pyroclastic rocks (Bennett 1978). Recent mapping by Jensen (1990) along the south limb of the Quirke Lake Syncline in the Pecors–Whiskey lakes area has both confirmed the presence and extended the known distribution of Huronian volcanic rocks in that area, as originally proposed by Robertson (1971).

The local development of paleosols on the Huronian volcanic rocks of the Quirke Lake Syncline has been observed by Bennett (1978), Robertson (1981), Roscoe (1981) and Meyer (1981). This is a sub–Matinenda Formation paleosol equivalent to that described by Bennett (1978) and Bennett et al. (1990) in the Otter–Haughton townships area. There is no evidence of volcanism in the Quirke Lake Syncline either during or after the deposition of the Matinenda

Formation. The presence of thin, but widespread, beds of locally radioactive, pyritic, quartz-pebble conglomerate, and arkose and grit, lying directly below or near the base of the Thessalon Formation in most areas, indicates that a period of erosion preceded the eruption of the first Thessalon flows, but followed the deposition of the Livingstone Creek Formation.

The association of the Livingstone Creek Formation with Huronian volcanic rocks (along with the latter's associated arkose and radioactive quartz-pebble conglomerate units) occurs over widely separated areas between the Quirke Lake Syncline and Sault Ste. Marie. A regional control on the deposition of these units is inferred. Some of these areas are adjacent to major thrust faults, which probably occupy the position of earlier normal faults (Zolnai et al. 1984). The presence of unconformities above and below the Thessalon Formation, in concert with the present distribution and thickness variation of the Thessalon and Livingstone Creek formations, may result from pre–Matinenda Formation erosion rather than original deposition.

EASTERN VOLCANIC ROCKS OF THE PENOKEAN FOLD BELT

The volcanic rocks of the Huronian Supergroup in the Sudbury–Spragge area have been given 4 lithostratigraphic, formation names. In the Sudbury area, 3 of these formations, in ascending order, are the Elsie Mountain, the Stobie and the Copper Cliff formations (Card 1978a). The Salmay Lake Formation, which lies at the base of the Huronian sequence in the Massey area (Robertson 1970), is the fourth, mainly volcanic formation in the eastern Penokean Fold Belt.

The Elsie Mountain Formation has a maximum preserved thickness of about 1000 m. The Stobie Formation is up to 1500 m, the Copper Cliff Formation 760 m, and the Salmay Lake Formation 450 m thick.

Elsie Mountain Formation and Stobie Formation

The volcanic rocks of the Elsie Mountain and Stobie formations are lithologically similar; the latter is distinguished

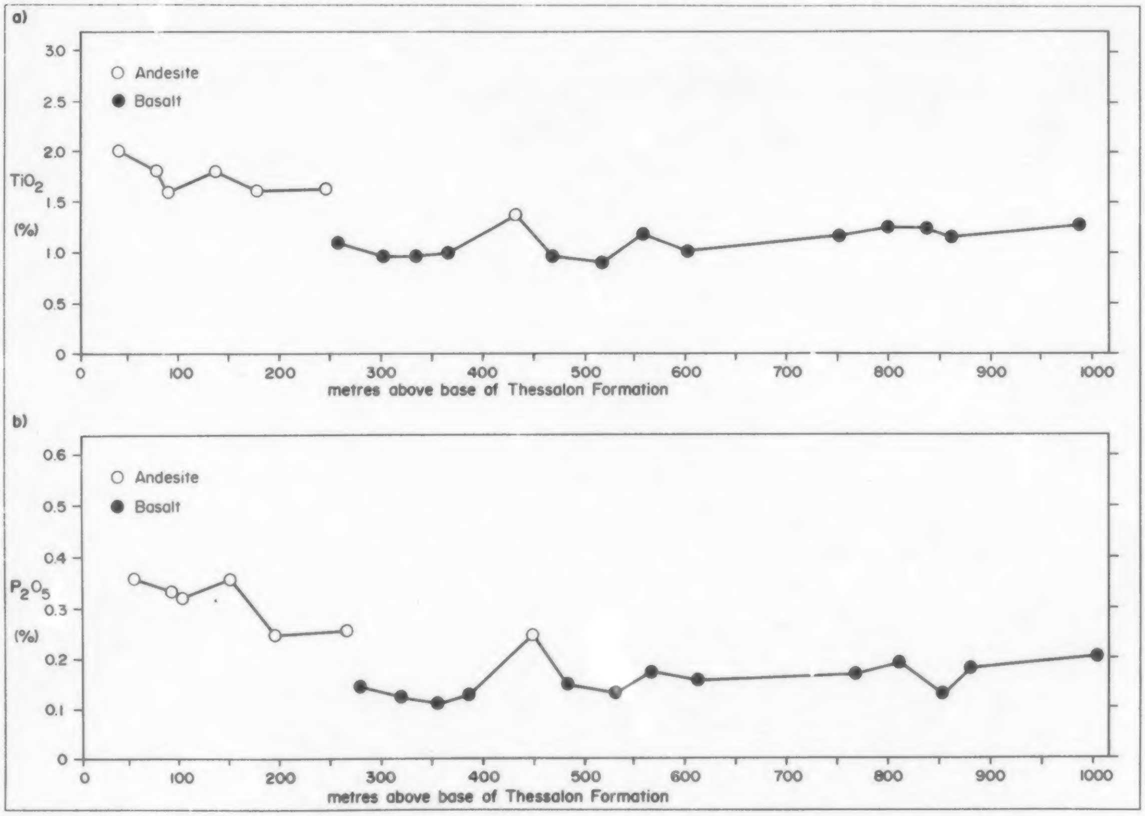


Figure 14.8. Variation of **a)** TiO_2 and **b)** P_2O_5 with stratigraphic height for the Thessalon Formation, near the Sault Ste. Marie area (*see* Figure 14.5 for orientation).

from the former by a higher proportion of intercalated sedimentary rocks. The volcanic units of both formations consist of dark green to black, foliate to massive, metamorphosed mafic flows, which dip steeply and face uniformly to the south. Pillow structures are common throughout the sequence (Figure 14.11). Individual flow thicknesses decrease with increasing stratigraphic height (Card 1978a; Innes 1977). A few amygdaloidal metadacitic flows exhibiting distinctive "eye-and-eyebrow" structures (Figure 14.12) occur within the Stobie Formation near Sudbury (Dressler and Muir 1986). Rhyolitic flows, up to approximately 5 m thick, are intercalated with mafic flows and sedimentary rocks in Falconbridge Township, northeast of Sudbury (Dressler 1987).

Mineral assemblages observed in these volcanic rocks of the Sudbury area are indicative of greenschist- to amphibolite-facies metamorphic rank (Card 1964; Fleet et al. 1987). Innes (1977) classified the Huronian volcanic rocks in the eastern part of the Penokean Fold Belt as subalkalic, slightly potassic tholeiite.

The middle and upper parts of the Stobie Formation contain a high proportion of intercalated wacke, siltstone, arenite and grit. Many of these clastic units are sulphide rich (Innes 1972, 1977) and some of them, although generally less than 3 m thick, can be traced laterally for up to 24 km. Some of the aluminous mudstones are rich in sericitized staurolite (Figure 14.13), muscovite and garnet (Card 1978a).

Copper Cliff Formation

The Copper Cliff Formation conformably overlies the Stobie Formation and contains intercalations of felsic and mafic flows, minor felsic pyroclastic rocks and metasedimentary rocks (Innes 1977). The rhyolite of the Copper Cliff Formation has been referred to as granite (Phemister 1956), arkose (Coleman 1905a, 1905b) and volcanics (Burrows

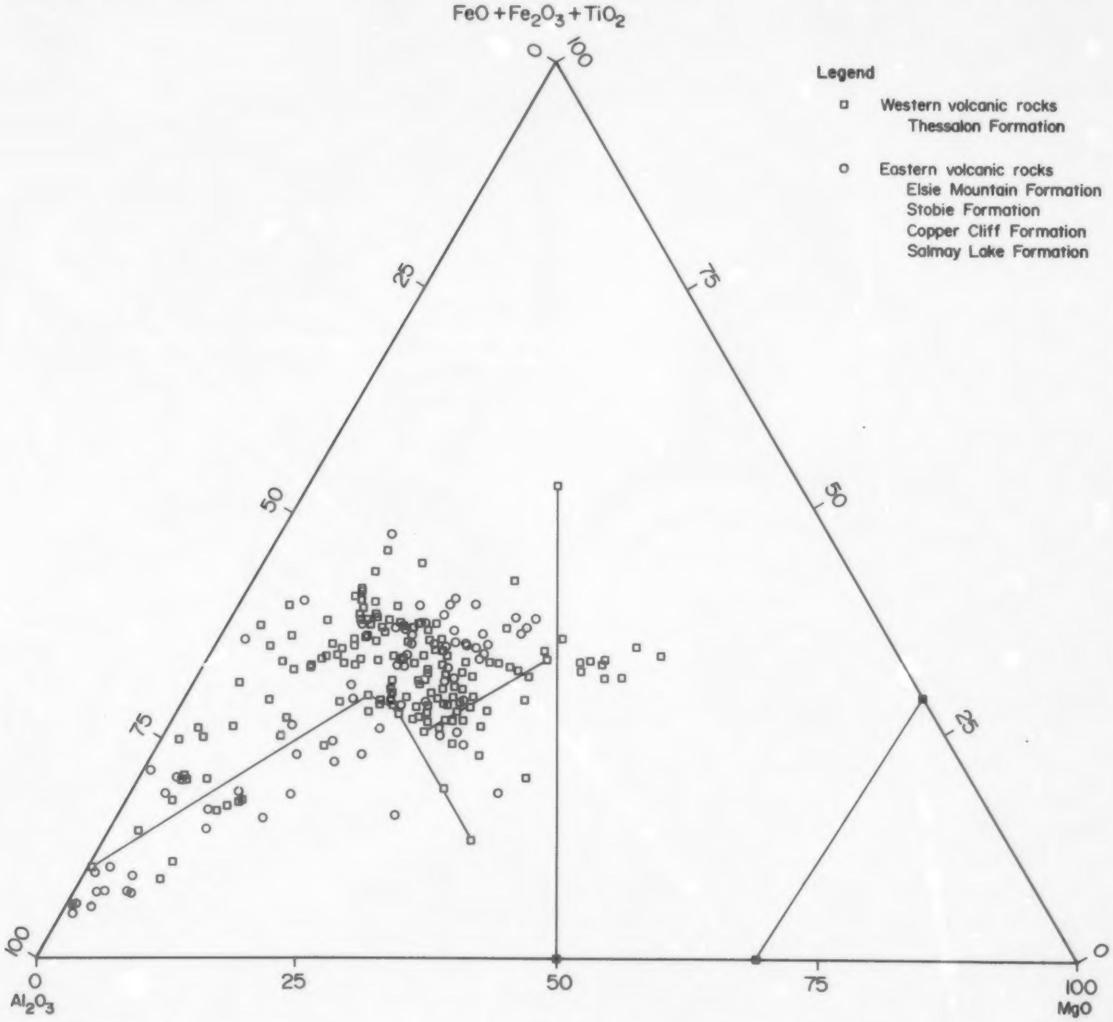


Figure 14.9. MgO–Al_2O_3–FeO+Fe_2O_3+TiO_2 cation plot (*modified from* Jensen 1976) of the Eastern and Western volcanic rocks, Huronian Supergroup.

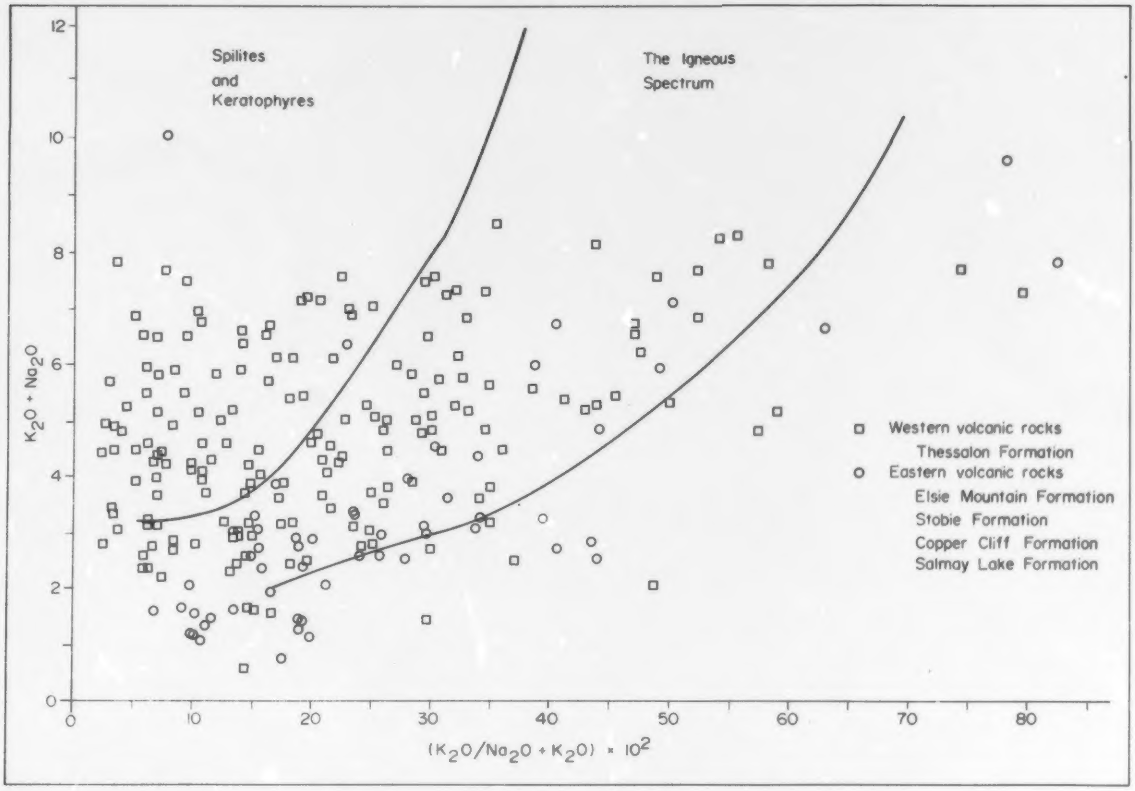


Figure 14.10. Igneous spectrum plot (*modified from* Hughes 1972) showing the sodic nature of many metavolcanic rocks of the Western metavolcanic rocks, indicating spilitization, in comparison to the Eastern volcanic rocks.

Figure 14.11. Amygdaloidal pillowed metabasalt, Stobie Formation, Falconbridge Township.

and Rickaby 1930). The rhyolite has yielded a U-Pb zircon age of 2450^{+25}_{-10}Ma (Krogh et al. 1984).

The felsic pyroclastic rocks and flows of the Copper Cliff Formation are medium- to fine-grained rhyolite and dacite (Card 1978a). Analyses of the felsic rocks indicate compositions including tholeiitic, potassic rhyolite and dacite (Card 1978a).

Salmay Lake Formation

The Salmay Lake Formation consists of massive, porphyritic and amygdaloidal basaltic and andesitic flows with minor pillowed basalt and rhyodacite. These rocks are lithologically similar to those of the Elsie Mountain Formation (Card 1978a). Interbedded wacke and quartz-pebble conglomerate (Robertson and Siemiatkowska 1972; Robertson 1976) resemble parts of the Matinenda Formation.

A 350 m thick sequence of steeply dipping, south-facing, amygdaloidal, pillowed metavolcanics in the Spragge area may be stratigraphically equivalent to the Salmay Lake Formation's volcanic rocks. The base of the section is fault-bounded; this sequence is conformably overlain by interstratified mudstone and arenite with minor pillowed, amygdaloidal and massive metabasaltic flows (Robertson 1970; Innes 1977).

SUMMARY—VOLCANIC ROCKS OF THE HURONIAN SUPERGROUP

Card (1978a) noted the unusual association of volcanic rocks with turbiditic wacke and well-sorted arkose in the Sudbury area. He suggested that the volcanic rocks were

Figure 14.12. "Eye-and-eyebrow" structures in a metadacite flow within the Stobie Formation, Sudbury area. The coin in the upper right corner is 2.5 cm in diameter.

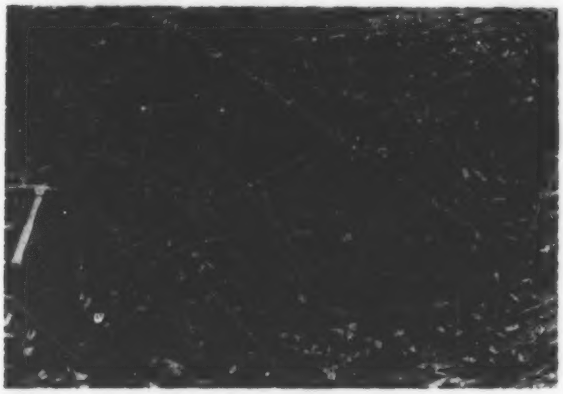


Figure 14.13. Staurolite (sericitized) in metapelitic rocks of the Huronian Supergroup, just west of Stobie Mine, McKim Township (*from* Dressler 1984).

derived from fault-controlled vents situated along the edge of a depositional basin and that the sandstones were derived from the older granitic terranes to the north. The turbidites, according to Card (1978a), may represent mixed detritus derived from both the volcanic and rejuvenated basement lying in a marginal position to the depositional basin.

Many contrasting features exist between the volcanic rocks of the western Penokean Fold Belt and those in the Sudbury area. Lithostratigraphic sequences of those 2 regions are also dissimilar in several aspects. For example:

1. In the Quirke Lake Syncline, the Matinenda Formation disconformably overlies the Huronian volcanic rocks, whereas in the Sudbury area there is much intercalation of Matinenda Formation, or "Matinenda-type" metasedimentary rocks with the metavolcanic rocks. If these "Matinenda-type" metasedimentary rocks are stratigraphically equivalent to the Matinenda Formation of the Elliot Lake area, then the volcanic rocks of the Sudbury area may be younger than the Huronian volcanic rocks within the Quirke Lake Syncline.
2. There have been no reports of paleosols within or upon the volcanic rocks in the Sudbury area (Eastern volcanic rocks) in contrast to their being present within the Huronian volcanic rocks in the Elliot Lake area (Western volcanic rocks).
3. The rhyolites of the western Penokean Fold Belt (Western volcanic rocks) are minor and restricted to the lower parts of the stratigraphy, whereas in the Sudbury area, the thick rhyolitic accumulations are high in the stratigraphy.
4. The metasedimentary rocks of the Thessalon Formation (western Penokean Fold Belt) are concentrated mainly in the lower part of the formation; in contrast to this, metasedimentary rocks are mainly found interbedded in the upper parts of the mafic volcanic sequences in the Sudbury area (eastern Penokean Fold Belt).
5. Andesites, mugearites and hawaiites have not been reported in the Sudbury area, but are present in the Thessalon Formation and stratigraphic equivalents in the Quirke Lake Syncline. The more sodic nature of the rocks of the Thessalon Formation probably reflects spilitic alteration, but the higher titanium and phosphorus contents may be primary.
6. Pillow structures and intercalated turbidites are common in the volcanic rocks of the Sudbury area and indicate subaqueous volcanism. Northeast of Sudbury, Dressler (1987) noted amygdaloidal mafic flows and minor rhyolites intercalated with conglomerate, indicating a state of shallow-marine to subaerial volcanism. However, a few pillowed, basaltic flows (*see* Figure 14.11) have also been reported in the same area. Nonetheless, features indicating subaqueous volcanism are not common in the Thessalon Formation.

Many of these differences between the western and eastern volcanic assemblages of the Huronian Supergroup may be ascribed to contrasting tectonic settings, to be described in a later section of this chapter.

Sedimentary Units of the Upper Elliot Lake Group

MATINENDA FORMATION

The Matinenda Formation (Roscoe 1957a, 1957b) hosts the uranium deposits of the Elliot Lake area (*see* Economic Geology).

The formation is up to 180 m thick in the Elliot Lake area (Robertson 1981). In the Thessalon and Sault Ste. Marie areas, the Matinenda Formation is probably less than 50 m thick (Bennett 1978), whereas just west of Sudbury it is as much as 600 m thick but thins rapidly eastward and becomes intercalated with the mainly metavolcanic rocks of the Stobie Formation and the mudstones of the McKim Formation (Card 1978a). Near the city of Sudbury, and to the east, the formation dies out (Dressler 1987).

The predominant rock types of the Matinenda Formation are medium-grained to coarse-grained, poorly sorted subarkose; arkose (Figure 14.14); and beds of pyritic and (in part) uraniferous, quartz-pebble conglomerate (Pienaar 1963; Roscoe 1969). The subarkoses and arkoses consist of poorly sorted quartz and feldspar grains set in a subordinate matrix of sericite and comminuted rock fragments. The ratio of potassic to sodic feldspar is approximately 8:1. Minor constituents are pyrite, calcite, chlorite, zircon and, rarely, leucoxene-coated iron oxides and monazite (Robertson 1968). Kimberley et al. (1980) reported that the uraniferous conglomerates contain almost no magnetite or ilmenite and have very high potash:soda ratios.

Roscoe (1969) proposed a threefold subdivision of the Matinenda Formation in the Quirke Lake Syncline of the Elliot Lake area. A lower, coarse-grained arkose, together with a uraniferous, pyritic, quartz-pebble conglomerate, forms a unit along the south limb of the syncline called the Ryan Member. This is overlain by a finer-grained subarkose and a polymictic conglomerate which together are called the Stinson Member; it also occurs along the syncline's south limb. Along the northern limb of the syncline, a second coarse-grained arkose accompanied by a uraniferous conglomerate—the Manfred Member—overlies the northern part of the Stinson Member. Later Roscoe (1981) added the Keelor Member to the top of the Matinenda Formation northwest of Elliot Lake. It consists of arkose, uraniferous and pyritic conglomerate, and conglomeratic arkose overlying the "Dollyberry volcanics" and probably the Stinson Member.

Scour-and-fill structures, trough and planar cross-bedding, and ripple marks are common in the arkosic units of the Matinenda Formation (Figure 14.15). Well-sorted, quartz-pebble conglomerate beds (Figure 14.16), containing well-rounded pebbles and cobbles of Archean-derived (?) quartz and chert, and pebbly subarkose units are commonly interbedded with the coarse-grained subarkose units, but are more common near the base of the formation. Various amounts of sericite give a green, apple-green or greenish yellow colour to the poorly sorted arkoses (Pienaar 1963; Robertson 1968).

Paleocurrent studies in the Elliot Lake area by McDowell (1957), Pienaar (1963), and Fralick and Miall (1989) showed that the Matinenda Formation was deposited from shallow, braided streams flowing down a south-dipping paleoslope which underwent tilting to the southeast during deposition.

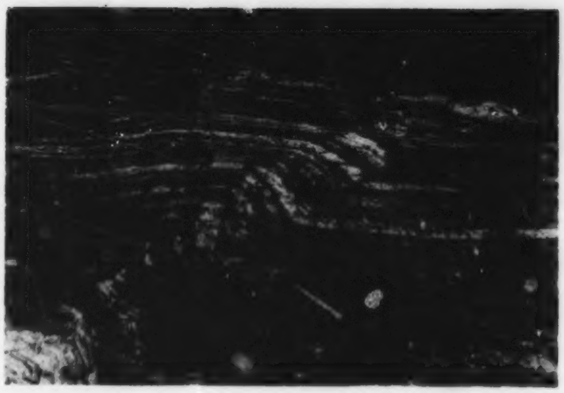


Figure 14.14. Bedded, kink-folded arkose of the Matinenda Formation, Hyman Township (*from* Dressler 1984).

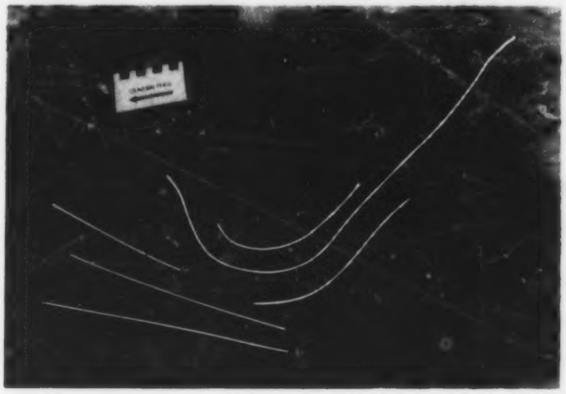


Figure 14.15. Trough cross-bedding in fine-grained subarkose of the Matinenda Formation, Highway 129, near Thessalon.

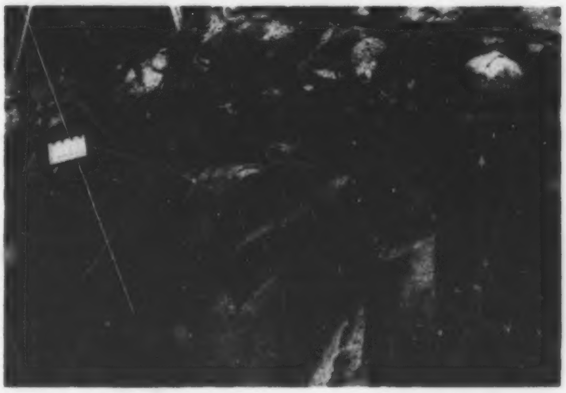


Figure 14.16. Pyritic quartz-pebble conglomerate of the Matinenda Formation in Otter Township. The disconformity (*see* arrow), overlies the paleosols of the Livingstone Creek Formation.

McKIM FORMATION

The McKim Formation (Robertson et al. 1969a, 1969b)—the "McKim greywacke" of Coleman (1914)—is the uppermost formation of the Elliot Lake Group. In the Sudbury area, it conformably overlies, and is locally intercalated with, volcanic rocks in the lower part of the Huronian Supergroup and the clastic sedimentary rocks of the Matinenda Formation. It is absent on the north limb of the Quirke Lake Syncline and in the Thessalon–Sault Ste. Marie area. In general, the formation extends from the Blind

River area to the Grenville Front Tectonic Zone. The thickness of the McKim Formation increases towards the south and east from the Elliot Lake area (Parviainen 1973). Robertson (1968) gives a thickness of up to 115 m for the McKim Formation on the south limb of the Quirke Lake Syncline. It is thickest in the Sudbury area, where it is up to 2400 m thick.

The McKim Formation is composed of subarkosic wacke, mudstone, subarkose, lithwacke and litharenite. Graded beds, parallel laminations, ripple marks, ripple cross-laminations and Bouma cycles, indicative of deposition by submarine turbidity currents, were reported by Parviainen (1973). Laminated mudstone and siltstone are prominent in the western part of the Sudbury area (Card 1978a). Card (1978a) also noted a change from predominantly laminated siltstone in the western end of the McKim Formation, to predominantly wacke in the eastern end, indicating a change from distal to proximal facies and possibly locating a source for the McKim Formation sedimentary rocks in the east. Intraformational breccias, slump folds, clastic dikes and ball-and-pillow structures are evidence for syndepositional tectonic activity (Card 1978a). In the Elliot Lake area, the McKim Formation represents a marine transgression gradually covering the coarse-grained, clastic sedimentary rocks of the Matinenda Formation (Fralick and Miall 1989; Card 1978a).

HOUGH LAKE GROUP

The Hough Lake Group (Roscoe 1969) is the lowest of the 3 Huronian groups which display cyclic deposition of conglomerate (Ramsay Lake Formation); mudstone, siltstone and turbidite (Pecors Formation); and cross-bedded arenite (Mississagi Formation). Each cycle (i.e., group) is thought to represent a sequence of glaciogenic, marine and fluvial to shallow-marine deposition. The Aweres Formation in the Sault Ste. Marie area may be equivalent to parts of the Ramsay Lake and Mississagi formations.

The lower contact of the Hough Lake Group with the Elliot Lake Group is generally conformable, except in the Agnew Lake area (Card and Palonen 1976) where there is evidence of minor erosion. The upper contact of the Hough Lake Group is erosional in places, based on the presence of channels within the Mississagi Formation filled with Bruce Formation (Debicki 1990). The contact is also considered unconformable by Card (1965b, 1976b).

RAMSAY LAKE FORMATION

The Ramsay Lake Formation (Coleman 1905a, 1905b) is the lowermost formation of the Hough Lake Group (Pienaar 1963; Roscoe 1969). The formation extends from Aberdeen Township, in the western part of the Penokean Fold Belt, to Falconbridge Township, northeast of Sudbury (Dressler 1987). The formation was named for Ramsay Lake (former spelling for Ramsey Lake), in Sudbury Regional Municipality, and was an important marker horizon both in mapping the Sudbury area and in mineral-exploration drilling in the Elliot Lake area.

In the Elliot Lake area, the Ramsay Lake Formation is up to 30 m thick. In the Sudbury–Manitoulin area (Card 1978a), the formation ranges from 70 to 170 m in thickness; within this area, a notable exception occurs in McKim Township where the formation is approximately 380 to 550 m thick (Debicki 1990). It is absent from Falconbridge Township (Dressler 1987).

By far the most abundant rock type within the Ramsay Lake Formation is matrix-supported, polymictic conglomerate (Figure 14.17). The clast compositions in the lowermost few metres of the formation usually reflect the underlying rock types (Robertson 1968; Parviainen 1973). Minor mudstone, wacke and arenite are locally present.

On the north limb of the Quirke Lake Syncline, the Ramsay Lake Formation overlaps the Matinenda Formation and, in places, lies on Archean basement. The Ramsay Lake Formation bevels and incorporates material from the uraniferous conglomerate of the Matinenda Formation, this being evident in the Quirke and Denison mines (Robertson 1968); but, where observed in the mines, the contact between conglomerates of the Matinenda and the Ramsay Lake formations may be either sharp or gradational over several metres. The intimate mixing of these units implies that the sediments of the Matinenda Formation were not lithified when the conglomerate of the Ramsay Lake Formation was deposited. The presence of uranium mineralization in the zone of mixing shows that it was already present in the Matinenda Formation. Frarey (1977) reported no evidence of a significant disconformity at the base of the Hough Lake Group in the Sault Ste. Marie–Elliot Lake segment of the Huronian Supergroup, a conclusion supported by Fralick and Miall (1989). However, Card and Palonen (1976) reported some evidence of erosion at the base of the Hough Lake Group near Agnew Lake.

Pienaar (1963) and Card (1978a) interpreted the conglomerates of the Ramsay Lake Formation to be debris

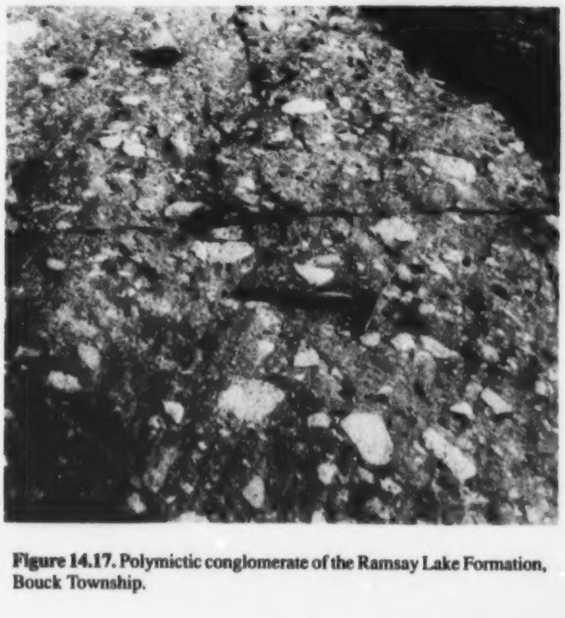


Figure 14.17. Polymictic conglomerate of the Ramsay Lake Formation, Bouck Township.

flows. Most of the other investigators advocated a glacial origin (Roscoe 1969; Robertson 1976). Parviainen (1973), for example stated that the depositional features of the Ramsay Lake Formation are most compatible with those of terrestrially deposited till. Parviainen noted the almost complete lack of internal stratification and structure as well as the lack of definitive indicators of composition, orientation, abundance and size of megaclasts which would otherwise have served as evidence of glacial origin from local source areas. The coarse-grained, sandy matrix is considered as evidence of the incorporation of unconsolidated material from the Matinenda Formation. The most recent study of the Ramsay Lake Formation (Fralick and Miall 1989) identifies a) an ice-proximal association of pebbly sandstone and mixtite; b) subaqueous gravity flows and ice-rafted deposits; and c) ice-proximal, fluvial outwash deposits.

PECORS FORMATION

The Pecors Formation conformably overlies the Ramsay Lake Formation and occurs in an area extending from about Quirke Lake to Sudbury. It was not identified between Thessalon and Sault Ste. Marie (Frarey 1977).

The Pecors Formation is 30 m thick at Quirke Lake (Robertson 1968) and is as much as 900 m thick south of the Murray Fault System in the Sudbury area (Card 1978a).

The formation consists of generally dark, bedded and laminated wacke, mudstone, siltstone and arenite (Roscoe 1969). Ripple marks, graded bedding, cross-laminations, parallel laminations, ball-and-pillow structures, clastic dikes and slump features are present. The basal part of the formation is commonly laminated, resembling a rhythmite, and, in places, has dropstones (Robertson 1968; Parviainen 1973). Partial Bouma sequences are common (Robertson 1976; Card 1978a).

The Pecors Formation, in the Sudbury–Espanola area, formed in deep water by deposition from turbidity currents (Card 1978a). At the north limb of the Quirke Lake Syncline, however, sedimentary features such as mud cracks indicate a shallow-water depositional environment. The presence of dropstones is evidence of a cool paleoclimate, and provides supporting evidence for the glacial origin of the underlying Ramsay Lake Formation.

MISSISSAGI FORMATION

The Mississagi Formation forms the uppermost division of the Hough Lake Group. It extends from near Sault Ste. Marie (Frarey 1977; Bennett and Sawiuk 1979) to as far east as the Cobalt Embayment (Card and Lumbers 1977).

In the Sault Ste. Marie area, the formation is about 1500 m thick (Frarey 1977); it ranges from 500 to 600 m in thickness in the Quirke Lake Syncline (Robertson 1968). Card (1978a) reported that 600 to 900 m of Mississagi Formation rocks occur north of the Murray Fault System, and that this increases in thickness to 1500 to 3000 m south of the Murray Fault System in the Sudbury area. However, according to Debicki (1990), the formation is about 3080 m thick north of the Murray Fault System in Falconbridge Township.

The Mississagi Formation consists mainly of moderately well-sorted, medium- to coarse-grained subarkose to arkose (Long 1976). Thin beds of dark chert-pebble and quartz-pebble conglomerate are common. Other rock types, ranging in composition from quartz arenite to wacke, make up minor parts of the formation (Card et al. 1977; Bennett 1982). Planar and trough cross-bedding, and ripple marks are common sedimentary structures (Figure 14.18). Sandstones of the lower half of the Mississagi Formation tend to be finer grained and less well sorted than those in the upper parts (Card et al. 1977; Bennett 1982).

Paleocurrent studies of the Mississagi Formation have been carried out by McDowell (1957), Pienaar (1963), Palonen (1971, 1973) and Long (1976). Long (1976) concluded that the Mississagi Formation was deposited from 2 major braided-stream systems. One system flowed south and east from the Sault Ste. Marie–Elliot Lake region to meet another flowing in a southwesterly direction from the Cobalt Embayment (Figure 14.19). However, it should be noted that the complex deformation of the Huronian Supergroup in the vicinity of the Sudbury Structure may make the paleocurrent reconstruction near Sudbury less certain. Debicki (1990) has shown a rather complex picture of paleocurrent directions for several Huronian formations near Sudbury. Long's (1976) regionally based fluvial model for Mississagi Formation sedimentation conflicts with a marine depositional model proposed by Palonen (1971, 1973) which is based on a more local study in the Lake Panache area, approximately 30 km southwest of Sudbury.

AWERES FORMATION

The Aweres Formation (McConnell 1927) is a sequence of conglomerates and sandstones, approximately 1700 m thick, which unconformably overlies the Thessalon Formation in the Sault Ste. Marie area. The precise lithostratigraphic position of the formation within the Huronian Supergroup is uncertain, but it is probably equivalent to the Mississagi Formation and perhaps, in part, to the Ramsay Lake Formation, as suggested by Frarey (1977).

The lower part of the Aweres Formation is characterized by both clast- and matrix-supported conglomerates with clasts derived mainly from the underlying volcanic rocks of the Thessalon Formation and sandstones of the Livingstone Creek Formation. Coarse-grained arkose predominates over conglomerate near the top of the sequence.

The matrix-supported conglomerates were probably deposited as debris flows forming alluvial fans since they are intercalated with beds of arkose and clast-supported polymictic conglomerate. Some beds of conglomerate display sieve texture, or lack fines and have interstices filled with calcite and quartz-albite mixtures.

QUIRKE LAKE GROUP

The Quirke Lake Group (Robertson et al. 1969a, 1969b); the "Quirke Group" of Roscoe (1957b)—previously termed the upper part of the "Bruce Series" (Collins 1914a, 1914b, 1925; Coleman 1914)—comprises 3 formations: the Bruce Formation, the Espanola Formation and the Serpent

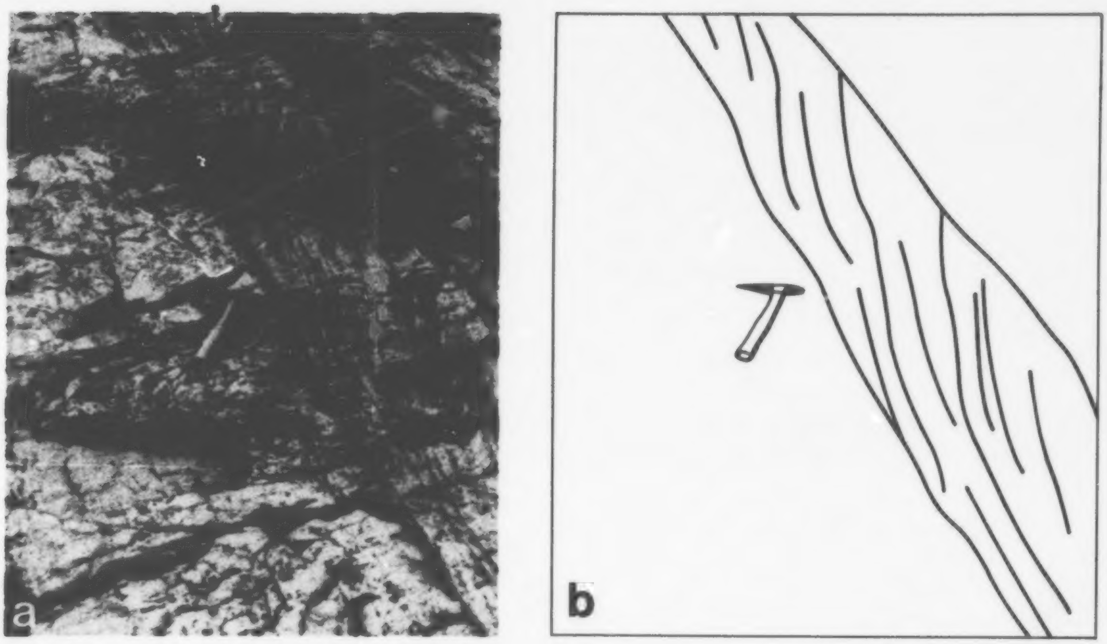


Figure 14.18. a) Planar cross-beds, Mississagi Formation, Garson Township (*from* Dressler 1984). The adjacent sketch **b)** emphasizes the crossbeds.

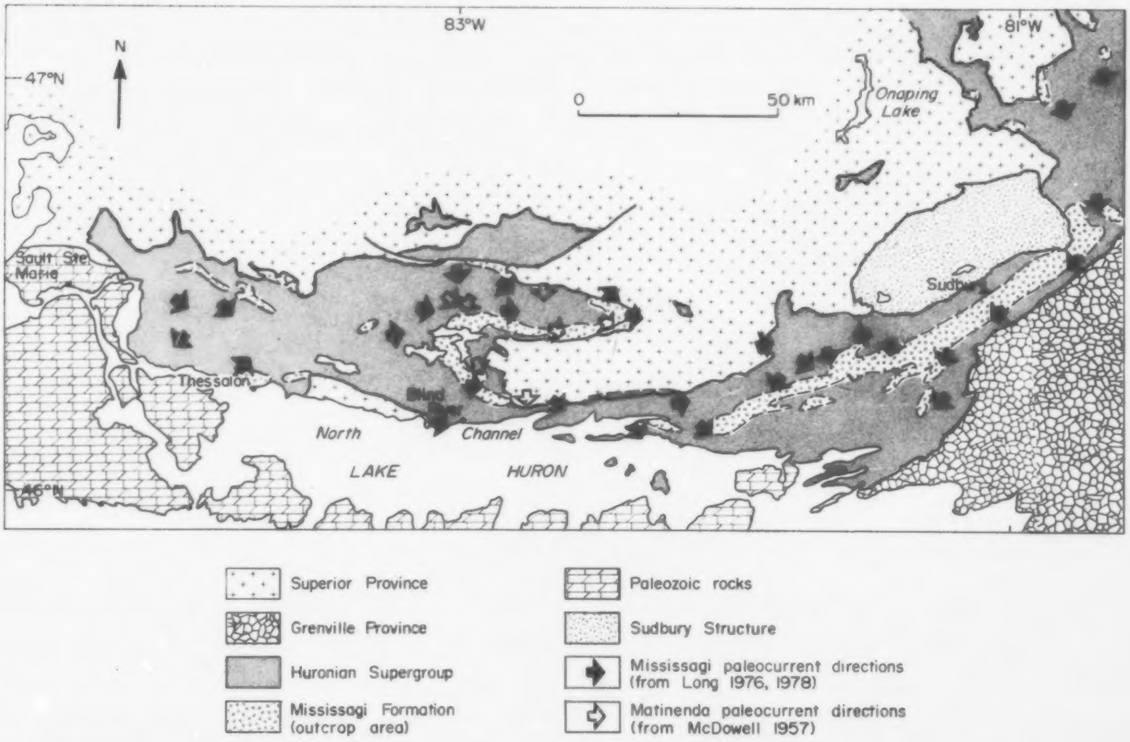

83°W
81°W
47°N
46°N
N
0
50 km
Onaping Lake
Sault Ste Marie
Thessalon
Blind
Sudbury
North
Channel
LAKE
HURON
Superior Province
Grenville Province
Huronian Supergroup
Mississagi Formation (outcrop area)
Paleozoic rocks
Sudbury Structure
Mississagi paleocurrent directions (from Long 1976, 1978)
Matinenda paleocurrent directions (from McDowell 1957)

Figure 14.19. Paleocurrent directions involved in the deposition of the Mississagi Formation. Directions have been determined from cross-stratification. Data *modified from* McDowell (1957) and Long (1976, 1978).

Formation. It extends from Sault Ste. Marie to Sudbury and to Yorston Lake in Seagram Township, northeast of Sudbury (Card et al. 1973); it may be present below the rocks of the Cobalt Group farther to the north of Seagram Township. The contact with the underlying arenites of the Mississagi Formation (Hough Lake Group) is conformable in places, but is considered to be unconformable by Card (1965a, 1976b) and Debicki (1990). Unconformable contacts with the wackes and matrix-supported conglomerates of the overlying Gowganda Formation of the Cobalt Group have been observed in several places. Zolnai et al. (1984) intimated structural reasoning to explain an upper, unconformable contact.

BRUCE FORMATION

The Bruce Formation ("Bruce conglomerate" of Collins 1914b) is known to occur in an area extending from Bruce Mines, at its western end, to the vicinity of Seagram Township (Card et al. 1973), 70 km northeast of Sudbury. It consists of polymictic conglomerate and pebbly wacke, and locally intercalated arkose, wacke and siltstone. The conglomerate above the Mississagi Formation was called the "Lower Slate conglomerate" by Logan and Murray (Logan 1863).

At Bruce Mine, the formation ranges from approximately 5 to 45 m in thickness; in the Sudbury–Manitoulin area, it is 60 to 400 m thick (Card 1978a); and near Falconbridge, it is approximately 250 m thick (Debicki 1990). In the Maple Mountain area (approximately 95 km northeast of Sudbury), Card et al. (1973) estimated the thickness of the Bruce Formation to be between 300 and 600 m. The contact with the underlying Mississagi Formation is locally disconformable (Collins 1925; Robertson 1961, 1962; Card 1978a).

The first detailed, comprehensive description of the "Bruce conglomerates" was given by Collins (1925) in a classic memoir entitled *The North Shore of Lake Huron*. He described the conglomerate as "a massive boulder conglomerate, consisting, typically, of subangular to round boulders of all sizes up to 3 feet in a dark grey matrix resembling greywacke". Since the time of Collins' investigations, the conglomerates and other rock types of the Bruce Formation have been shown by many studies to be homogeneous in general, but with some variations in detail. Matrix-supported, and minor clast-supported, conglomerate is the main component of the formation (Figure 14.20). The clasts are subangular to rounded, and commonly consist of whitish granite, Archean supracrustal rocks, quartzite, and, in places, metasedimentary rocks of Huronian formations older than the Bruce Formation. The matrix is a wacke characterized by interstitial pyrite and rounded to subrounded, glassy quartz grains up to 1.5 mm across. Collins (1925) stated that the matrix has experienced some "decay and washing-out of its finer clay-like components from the coarse part". This, according to Collins, distinguishes the "Bruce conglomerate" from other Huronian conglomerates. As much as 5% carbonate is present in the matrix of rocks in the upper parts of the formation.

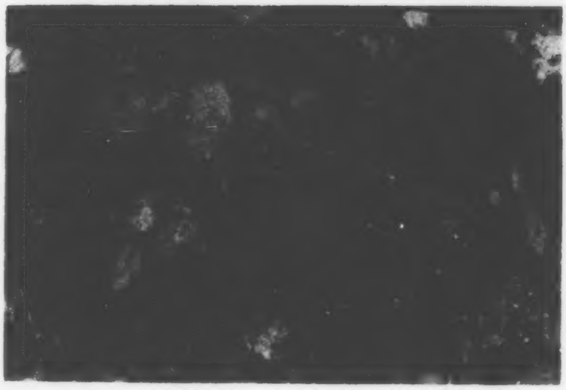


Figure 14.20. Polymictic conglomerate, Bruce Formation, Falconbridge Township.

Rock types within the Bruce Formation are similar to those of the other formations in the Huronian Supergroup. For example, arenites are similar to those in the underlying Mississagi Formation; wackes and siltstones are similar to those in the overlying Espanola Formation, especially where they contain carbonates.

The Bruce Formation has been interpreted as a tillite containing minor beds and lenses of glacially derived sandstone and siltstone. Ginn (1961) observed that pebbles appear to have been dropped into the siltstones before its lithification. This observation had also been made by Robertson (1964, 1971), who described clasts which had been dropped into rhythmite lenses just below the overlying Espanola Formation. According to Casshyap (1969), however, the diamictites of the formation were deposited by terrestrial, wet-base glaciers. However, Sims et al. (1981) interpreted the Bruce Formation as an accumulation of debris flows caused by normal faulting, subsidence and a sudden increase in paleoslope and water depth. In this chapter, we favour a glacial origin for the Bruce Formation. A debris-flow origin is considered unlikely as it does not explain the great variety of clasts in the conglomerates which, over large areal extents, overlie monotonous, thick arenite sequences of the Mississagi Formation. Debris-flow clasts should reflect more the lithology of the underlying rock units rather than what is actually exhibited by the clasts of the polymictic conglomerates of the Bruce Formation.

ESPANOLA FORMATION

The Espanola Formation (Collins 1914b) occurs from Sault Ste. Marie to northern DeMorest and Clary townships in the Maple Mountain area (Card et al. 1973), 70 km northeast of Sudbury. It is the earliest, significant carbonate-bearing formation of the Huronian Supergroup and, in places, can be lithostratigraphically subdivided into 3 parts: a lower carbonate member, a middle siltstone-wacke-arenite member, and a dolomitic member at the top. The lower carbonate member was formerly called the "Bruce limestone" (Winchell 1888). In the Maple Mountain area, the dolomitic member appears to be absent and instead a calcareous, silty wacke predominates at the top of the formation. Near Espanola, an upper cross-bedded quartz-arenite member is present (Card 1976a, 1978a). Near Sault Ste. Marie, the Espanola Formation is a few metres thick; in the Espanola–Lake Panache area it reaches approximately 600 m in thickness (Card 1978a); and in the Maple Mountain area it has an average thickness of about 40 m (Card et al. 1973).

Collins (1925) described the "Bruce limestone", the "Espanola greywacke" and the "Espanola limestone" as 3 formations of his "Bruce Series". In 1937, he combined these 3 units into the Espanola Formation (Collins 1936). The formation represents a well-defined marker horizon within the Huronian Supergroup lithostratigraphy.

All 3 members are commonly thinly bedded or laminated (Figure 14.21). The lower carbonate member grades into the wacke and mudstone of the middle member. This middle member commonly contains both minor interlaminated carbonate beds and carbonate disseminated throughout in various amounts. The "Espanola limestone" (Collins 1914b)—the commonly dolomitic upper member of the Espanola Formation—has gradational contacts with both the underlying middle member's wackes and the overlying Serpent Formation's arenites. However, disconformable contacts with the overlying Serpent Formation were also observed (Frarey 1977). Intraformational breccias and breccia dikes have been observed in the Espanola Formation. Other sedimentary features include mud cracks, ripple marks, flame structures and ball-and-pillow structures. Hoffman et al. (1980) described crustose and columnar carbonate stromatolites from the Espanola Formation at Quirke Lake. This bottom-dwelling biota from the photic zone, combined with the sedimentary features of the commonly fine, rhythmically interlayered limestones, wackes and mudstones, suggest deposition in quiet, relatively shallow water. Carbonate precipitation was intermittently interrupted by the influx of fine-grained detritus. Carbonate precipitation and benthic life-forms such as stromatolites are indicative of a moderate to warm climate during deposition, possibly reflecting a tidal-flat environment. Carbonate sedimentation at high latitudes (Williams 1975) and the association of detrital uraninite with intense chemical weathering (Maynard et al. 1991) suggested that any climatic modelling for the Huronian Supergroup is still uncertain. The sharp change of rock type from the Bruce Formation to the Espanola Formation has been interpreted as an indication of an abrupt change from a glacial to a moderately warm paleoclimate (Young 1973a). Such interpretations must now be viewed cautiously given the suggestions of Maynard et al. (1991) summarized above.

Skarns have developed in the carbonate rocks juxtaposed against Nipissing intrusive rocks. Scapolite is regionally developed in the Espanola Formation (Card 1978a).

SERPENT FORMATION

The Serpent Formation (Collins 1914b, 1925) extends from Echo Lake near Sault. Ste Marie, in the western Penokean Fold Belt, to Sudbury and DeMorest and Clary townships, 70 km northeast of Sudbury, in the Cobalt Embayment. It is possibly present farther north within the Cobalt Embayment beneath the rocks of the Cobalt Group.

Figure 14.21. Bedded, folded limestone of the Espanola Formation (photograph *courtesy of* W. Meyer).

The thickness of the formation ranges from 150 to 250 m at Bruce Mines (Frarey 1977), to 1500 m in the Sudbury–Espanola area (Card et al. 1977), 1000 to 1500 m in Falconbridge Township (Dressler 1987), and 600 m in the Maple Mountain area, northeast of Sudbury (Card et al. 1973). The variation in thickness is possibly attributable to erosion prior to the deposition of the Gowganda Formation, which locally unconformably overlies the Serpent Formation.

The Serpent Formation consists mainly of feldspathic quartz arenites and arkoses. These rocks are fine to medium grained and exhibit sedimentological features such as planar cross-bedding, festoon cross-bedding, fine lamination, ripple marks and mud cracks (Figure 14.22; Young et al. 1977; Young 1981a, 1981b). Conglomeratic rocks have been observed near the base of the formation in most occurrences and consist of thin lags, lenses, and interbeds of polymictic, clast- and matrix-supported conglomerate. Card et al. (1977) described a general upward increase in average grain size and degree of sorting within the formation. However, silty units have also been noted in the upper part of the formation. Silty limestone and carbonate-bearing units of siltstone and wacke have been observed by Robertson (1961, 1968) and Card et al. (1977).

Sedimentological features are suggestive of a subaerial to shallow-marine (Young et al. 1977; Young 1981a, 1981b) or possibly a distal braided-stream environment (Long 1976). Discontinuous, calcareous-siltstone lenses and beds represent sabkha deposits (Long 1976).

COBALT GROUP

The Cobalt Group (Miller 1913; Collins 1917)—the upper most sedimentary cycle of the Huronian Supergroup—comprises 4 formations: the Gowganda, Lorrain, Gordon Lake and Bar River formations. The group extends from Sault Ste. Marie through to Kirkland Lake in northeastern Ontario, and into northwestern Quebec. The Gordon Lake and Bar River formations, however, commonly have a more restricted areal extent. A small remnant of the Cobalt Group occurs about 70 km north of Sault Ste. Marie, suggesting that the areal extent of the group originally was much larger. At its thickest, near Lake Panache, southwest of Sudbury, the group is about 5800 m thick. Note that estimates of stratigraphic thickness undertaken south of the Murray Fault System are tentative because the extent of tectonic modification upon the stratigraphy of the Cobalt Group is unknown.

The contact with the underlying Serpent Formation of the Quirke Lake Group varies from disconformable to conformable. An angular unconformity exists northeast of Sudbury. Throughout much of the Quirke Lake Syncline, the Gowganda Formation lies upon the Espanola Formation. Fragments of rocks derived from the Espanola and Serpent formations have been observed in the basal portions of the Gowganda Formation (Collins 1925; Robertson 1963, 1968; Pienaar 1963). These observations just described, along with structural arguments advanced by Card et al. (1973), Card and Palonen (1976) and Zolnai et al. (1984) provide evidence of pre–Gowganda Formation erosion and folding.

GOWGANDA FORMATION

The Gowganda Formation (Collins 1917) occurs in a large area extending from Sault Ste. Marie to Sudbury (Penokean Fold Belt) and continuing (Cobalt Embayment) toward the Timmins–Kirkland Lake area of northeastern Ontario. It consists of distinct, diverse sequences of clast- and matrix-supported conglomerate, pebbly wacke, wacke, siltstone, mudstone and arenite. Its thickness ranges from 1070 m near Sault Ste. Marie; from 970 to 1150 m around Whitefish Falls, on the north shore of Lake Huron; and from 950 to 2700 m in the northern Wanapitei Lake area, near Sudbury. In the Maple Mountain area, the Gowganda Formation reaches a thickness of 1500 m; in the Temagami area, it attains approximately 700 m of thickness. Lateral and vertical facies changes, and intraformational unconformities are common. Wackes and fine-grained rocks are more common in the upper parts of the formation. It was this characteristic feature, occurring in the Cobalt Embayment northeast of Sudbury, which led to a subdivision of the Gowganda Formation into a lower conglomeratic unit—the Coleman Member—and an upper, finer-grained, silty wacke and mudstone unit—the Firstbrook Member.

The origin of the Gowganda Formation has been much discussed. Occurrences of till-like conglomerate, dropstones in wacke and siltstone, varved wacke and siltstone, striated clasts, and possibly a striated pavement occurring beneath the formation all point to a glacial, glacial-marine or glaciolacustrine origin of the formation (amongst others: Coleman 1905a, 1905b; Ovenshine 1965; Schenk 1965; Casshyap 1969; Lindsey 1971; Young and Nesbitt 1985; Legun 1981; Miall 1983, 1985; Mustard 1985; Chandler 1986). Striations on clasts have been observed, but it should be noted that they could be tectonic in origin (Robertson 1971); moreover, the identification of what is interpreted as striated pavement is not universally accepted. A debris-flow origin for the conglomerates and a deposition from turbidity currents for the finely laminated sediments have also been proposed. A combination of debris flow or turbidity current and a glacial origin represents an alternative hypothesis. Rosen (1985), for example, interpreted the Gowganda Formation near the Gowganda–Elk Lake area as being a product of both submarine gravity flows and ice-rafting of debris in a marine, ice-shelf environment. The commonly finer-grained upper parts of the formation have been interpreted by Junnila (1986)—reporting on work conducted in the Whitefish Falls area—to be subaqueous deposits of a delta which prograded to the southeast into a wave-dominated, shallow-marine environment. A deltaic, depositional

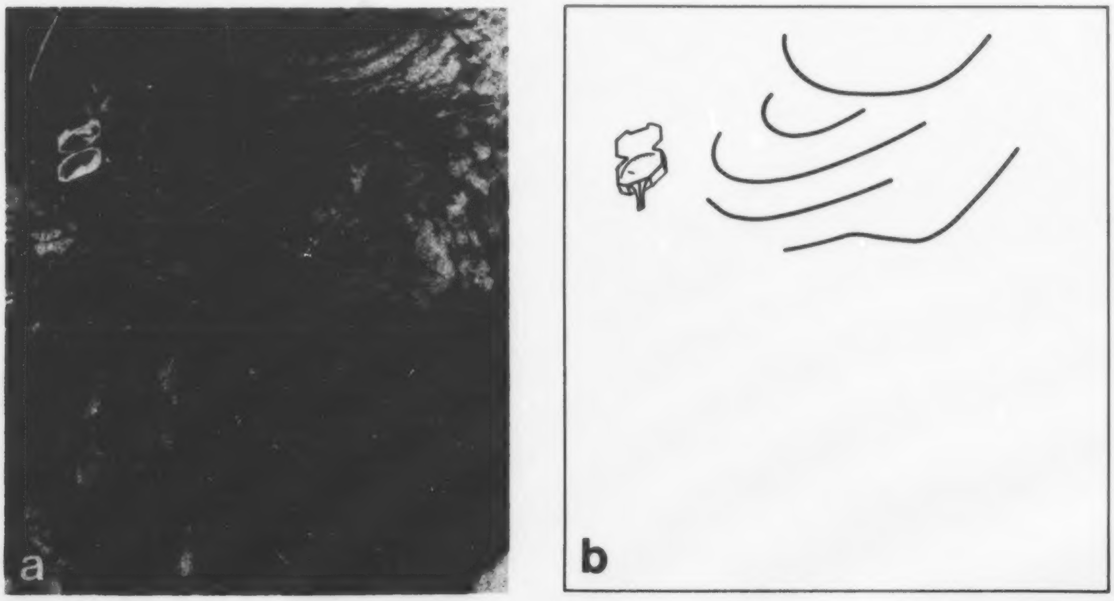


Figure 14.22. a) Trough cross-beds, Serpent Formation, Street Township (*from* Dressler 1984). The adjacent sketch **b)** emphasizes the cross-beds.

environment was also proposed for the Firstbrook Member, in the eastern Cobalt Embayment, by Rainbird (1986) who interpreted the lower, middle and upper parts of the Firstbrook Member to represent prodelta, delta-slope and delta-foreslope environments, respectively.

LORRAIN FORMATION

The Lorrain Formation (Miller and Knight 1905) underlies large areas within the Cobalt Embayment. It is known to occur from Sault Ste. Marie to Sudbury, in the Penokean Fold Belt, and in the Timmins–Kirkland Lake area. The formation consists mainly of arkose and quartz arenite and conformably overlies the Gowganda Formation. The contact is transitional and is commonly drawn at the base of the first major arkose unit. The formation is approximately 2400 to 2500 m thick near Sault Ste. Marie, 2500 m thick in the La Cloche Syncline southwest of Sudbury, up to 3300 m thick north of Wanapitei Lake, and 1500 to 2300 m thick in the Maple Mountain area of the Cobalt Embayment.

Several attempts have been made to subdivide the Lorrain Formation into subunits or members: in the Flack Lake area, 22 km northwest of Elliot Lake (Robertson 1971); in the La Cloche Lake area, 10 km south of Massey (Robertson 1976); and in the Maple Mountain area of the Cobalt Embayment (Card et al. 1973). For all these areas, a general upward increase in maturity-index values of the arenites is present. For example, the 7 members proposed by Card et al. (1973) have a general range in composition from matrix-rich, arkose and subarkose at the bottom of the formation, to greenish arkose and wacke in the middle of the formation, to a mature white quartz arenite at the top. Card et al. (1973) reported hematitic arenite occurring beneath both the upper, white quartz arenite and a micaceous, white quartz arenite. A jasper-pebble conglomerate member (Frarey 1977), characterized by bright red jasper pebbles set in a white quartz-arenite matrix, occurs in the Sault Ste. Marie–Bruce Mines area and has been used for decorative or craft purposes.

In the Mount Lake–Rawhide Lake area, the basal part of the hematitic sequence includes a horizon marked by thorium-bearing quartz-pebble bands containing monazite and zircon. The thorium-rich, uranium-poor, nonpyritic character of the unit is in marked contrast to the pyrite-bearing uranium mineralization found in the drab-coloured rocks of the lower parts of the Huronian Supergroup; in essence, it reflects the increased oxygen level of the atmosphere during the time of deposition of the upper sequences of the Huronian Supergroup (Frarey and Roscoe 1970).

A striking feature of the Lorrain Formation's upper quartz-arenite member north of Lake Huron is its content of secondary aluminous minerals in the matrix. Kyanite, andalusite and kaolinite occur in metamorphosed quartz arenite from the Whitefish Falls area. Diaspore and kaolinite occur in the rocks north of Bruce Mines (Chandler et al. 1969) and north of Elliot Lake (Wood 1970a, 1970b, 1973). Young (1973b) and Wood (1973) interpreted diaspore and kaolinite as early, postdepositional and premetamorphic minerals that result from *in situ* surficial alteration of feldspar in a hot, humid, and possibly tropical climate. In the La Cloche Lake–Killarney area, the above-listed aluminous minerals are metamorphosed to kyanite and andalusite (Robertson 1976; Card 1978a).

A wide range of primary structures have been observed. Amongst these are planar and trough cross-beds, ripple marks, convolute bedding, ball-and-pillow structures, low-angle scours, gravel lags and rare occurrences of mud chips. All these features occur in both shallow-marine and fluviatile environments, and it is therefore understandable that no consensus has been reached on the depositional environment of the Lorrain Formation. Wood (1973) and Frarey (1977), amongst others, advocated a fluviatile origin; Frarey and Roscoe (1970) upheld the view of a fluviatile–fluviatile-deltaic origin. Pettijohn (1970) advanced a marine-environment model, and Card (1976a) proposed a near-shore coastal-shelf environment in which the sediments of the Lorrain Formation were laid down during cyclically repeated marine transgressions and regressions. Rice (1987, 1988, 1991) interpreted the basal conglomerate of the formation in Lorrain Township "to represent debris flow and stream flow deposition in a proximal to intermediate position on a basin marginal alluvial fan(s)" and, in general, favoured a marine, siliciclastic shelf rather than a fluvial, sandy braid-plain paleoenvironment for the formation. He proposed an episodic, marine transgressive regime with several stillstands to account for his sedimentological observations. During the inundations, at least partial and possibly complete reworking of the braid-plain arenites occurred. This episodic transgression model appears to account for many of the controversial observations and interpretations of the paleoenvironment of the Lorrain Formation.

GORDON LAKE FORMATION

The Gordon Lake Formation (Frarey 1967) was informally known as the "Banded Cherty Quartzite Formation" (Collins 1925). It occurs in the Bruce Mines–Lake George area (Frarey 1977); in the Flack Lake area, northwest of Elliot Lake; southwest of Sudbury, in the La Cloche mountain area; and in the Cobalt Embayment. Nowhere is it very extensive. Contacts with both the underlying Lorrain Formation and the overlying Bar River Formation are gradational and conformable.

The formation is approximately 300 to 360 m thick in the Blind River area (Frarey 1977), approximately 700 m thick in the Sudbury–Espanola area (Card et al. 1977), and 540 to 600 m thick in the Maple Mountain area (Card et al. 1973); near Sault Ste. Marie, it is approximately 300 m thick.

The Gordon Lake Formation consists of well-bedded, variegated mudstone and siltstone; chert; and minor, fine-grained quartz sandstone. Robertson (1986) described 3 members from the Flack Lake area: a lower member consisting of reddish arenite and siltstone with chert, anhydrite and gypsum nodules (Chandler 1988); a middle member composed of green siltstone and mudstone with minor arenite; and an upper member made up of reddish siltstone, mudstone and chert (Eisbacher and Bielenstein 1969). In the Maple Mountain area, Card et al. (1973) also noted 3 members, all consisting of variously interbedded siltstones and arenites. Robertson (1976) has noted that the middle member of the formation in the Flack Lake area, and the formation in the La Cloche Syncline has a slight but distinctive magnetic expression.

Sedimentological features, such as current ripples, small-scale cross-bedding and desiccation cracks, and the local presence of anhydrite and gypsum, indicate deposition in a nearshore, marine, stable, shallow-water and low-energy, tidal-flat or lagoonal environment.

BAR RIVER FORMATION

The Bar River Formation (Frarey 1967), previously known as the "upper white quartzite" (Collins 1925), conformably overlies the Gordon Lake Formation with which it shares a similar areal distribution. Near Gordon Lake and Diamond Lake, northwest of Bruce Mines, and also near Flack Lake, northwest of Elliot Lake, the Bar River Formation has a preserved thickness of approximately 300 m. In the Sudbury–Espanola area of the Penokean Fold Belt (Card 1978a), the Bar River Formation is about 900 m thick whereas, in the Maple Mountain area (Card et al. 1973) of the Cobalt Embayment, it is 400 to 600 m thick. Near Sault Ste. Marie the formation is about 310 m thick.

The formation is characterized by mature quartz arenite; ferruginous arenite and siltstone have also been observed. Sedimentological features include massive bedding, cross-beds, ripple marks and, locally within finer-grained rocks, mud cracks and desiccation features. Some features were interpreted as possible worm casts, but are now generally regarded as infillings of desiccation features (*see* Robertson 1976 for discussion).

Wright and Rust (1985) interpreted the Bar River Formation's paleoenvironment as being shallow marine, meso- to macro-tidal whereas Robertson (1986) considered the formation in the Flack Lake area to represent either a beach or eolian deposit.

PALEOSOLS WITHIN THE HURONIAN SUPERGROUP

Paleosols have been observed at several locations in areas underlain by rocks of the Huronian Supergroup. They provide insight into the paleogeographic, paleoclimatic and structural development of the supergroup and have been studied by several researchers. Paleosols underlying the Matinenda Formation, in the Elliot Lake area, were described by Roscoe (1957a, 1957b), Pienaar (1963) and Robertson (1963, 1968, 1970). Collins (1925) described a paleosol in the Thessalon area where Archean granitic rocks change gradationally into overlying granite-boulder conglomerates of the Livingstone Creek Formation.

The paleosols are best developed in both Huronian Supergroup and Archean-basement rocks underlying the Matinenda Formation in the western part of the Penokean Fold Belt. They have not been reported east of the Quirke Lake Syncline or in rocks stratigraphically above the Elliot Lake Group. Huronian paleosols satisfy most criteria for the identification of paleosols as outlined by Grandstaff et al. (1986). On mafic volcanic rocks, paleosol can be recognized by the distinct apple-green to yellowish colour which generally grades over a few centimetres to several metres to dark-coloured parent rock. Locally the sericite-rich rock passes to an underlying black, fine-grained, massive, chlorite-rich eluvial zone with minor sericite (Figure 14.23).

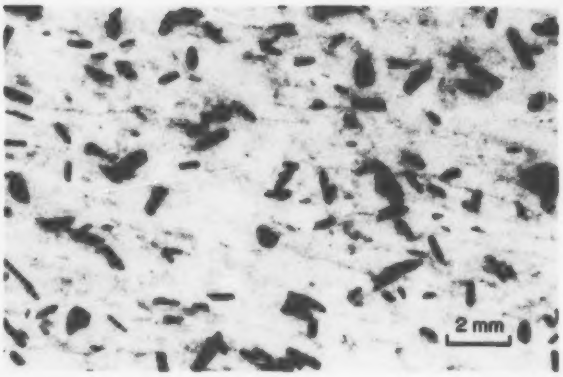


Figure 14.23. Photomicrograph of sub–Matinenda Formation paleosol of Huronian metavolcanic rock showing pseudomorphs of leucoxene after ilmenite in sericite groundmass, Elliot Lake area.

Most studies of Huronian paleosols have shown that they are characterized by a pronounced loss of soda (Na_2O), lime (CaO) and magnesia (MgO). There is a large increase in potash (K_2O)(Pienaar 1963; Robertson 1963, 1968, 1970). The amounts of Fe and Mn are generally depleted, but Gay and Grandstaff (1980) noted an increase in ferric iron and Mn in the upper part of a paleosol from the Pronto Mine area (Long Township) which they concluded to be an indication of the presence of free oxygen in the early Huronian atmosphere. However, Grandstaff (1980) estimated that the partial pressure of oxygen in the Huronian atmosphere was less than 1% of the present level. The enrichment in potash (as sericite) in Huronian paleosols is considered to be due largely to diagenetic and metamorphic processes which may mask the environmental and hydrologic conditions operative during paleosol development (Gay and Grandstaff 1980; G-Farrow and Mossman 1988).

The Huronian Atmosphere

The shallow-marine to subaerial, clastic units of the lower 3 Huronian groups, with few exceptions, lack red colours. Roscoe (1969, 1973) and Frarey and Roscoe (1970) interpreted this as being due to an oxygen-deficient atmosphere which resulted in a reduction of Fe_2O_3 to FeO during the deposition of rocks of the 3 lower Huronian groups. This interpretation may be substantiated by the presence of abundant pyrite and detrital uraninite in these Huronian rocks. However, some authors suggested that detrital uraninite may be due to intense chemical weathering during production of quartz-rich sediments (Maynard et al. 1991). Further evidence for an oxygen-deficient atmosphere during weathering, transport and deposition of early Huronian sediments is the loss of iron in paleosols. The loss of iron could be

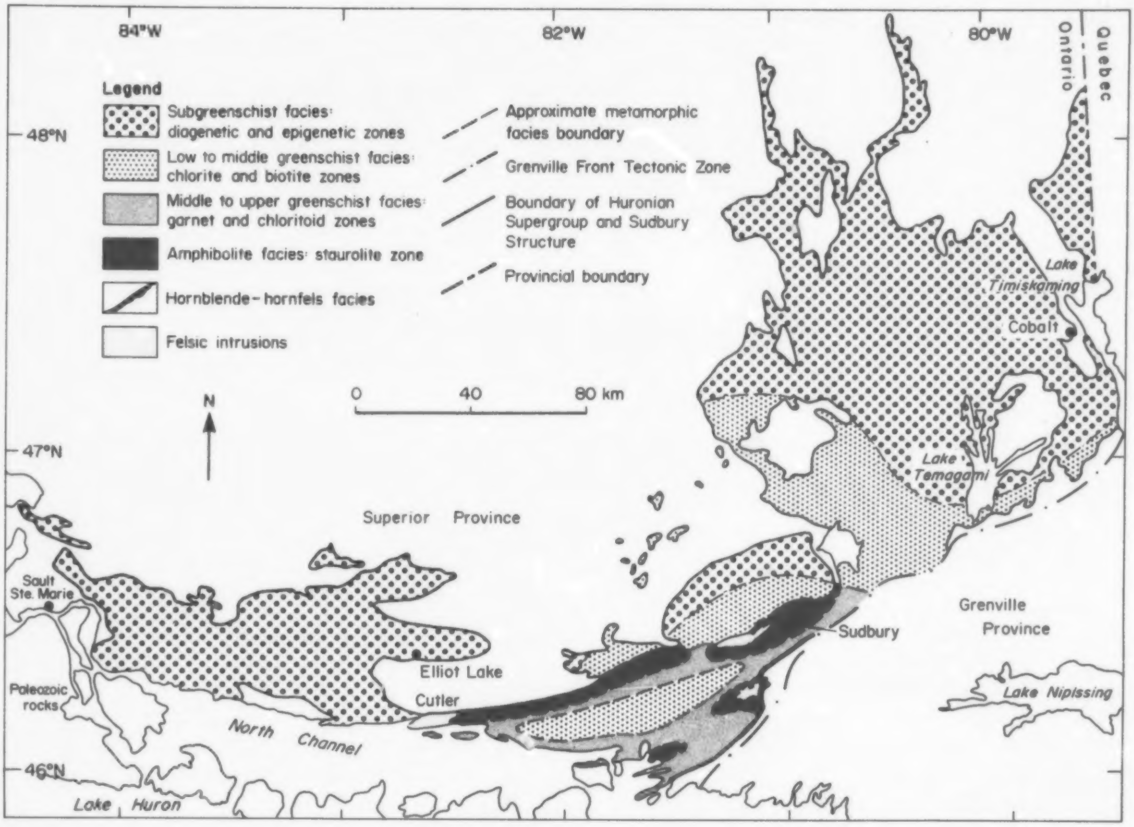

84°W
82°W
80°W
Ontario
Quebec
48°N
47°N
46°N
Legend
Subgreenschist facies: diagenetic and epigenetic zones
Low to middle greenschist facies: chlorite and biotite zones
Middle to upper greenschist facies: garnet and chloritoid zones
Amphibolite facies: staurolite zone
Hornblende-hornfels facies
Felsic intrusions
Approximate metamorphic facies boundary
Grenville Front Tectonic Zone
Boundary of Huronian Supergroup and Sudbury Structure
Provincial boundary
N
0
40
80 km
Superior Province
Lake Timiskaming
Cobalt
Lake Temagami
Sault Ste. Marie
Paleozoic rocks
Elliot Lake
Cutler
North Channel
Lake Huron
Sudbury
Grenville Province
Lake Nipissing

Figure 14.24. Distribution of metamorphic facies for the Huronian Supergroup, eastern Southern Province (*modified from* Card 1978b).

entirely due to climatic factors; for example, iron-poor calcrete and bauxite develop in the modern tropical environment. Red and reddish rocks, which mark the presence of an oxidizing atmosphere, make an appearance with the Gowganda Formation of the Cobalt Group and are widespread in parts of the Lorrain and Gordon Lake formations of that group. Lorrain Formation pebble beds containing monazite, but no uraninite, and hematite, rather than pyrite, are also suggestive of an atmosphere containing more oxygen than the atmosphere of lower "Huronian time".

Not all writers have agreed that the early Huronian atmosphere was reductive. Dimroth and Kimberley (1976) have argued from the distribution of sedimentary carbon, sulphur, uranium and iron that the Huronian atmosphere may not have been as oxygen deficient as had been proposed. Geochemical studies of paleosol profiles by Grandstaff (1980) and Gay and Grandstaff (1980) found that one of the paleosols studied showed evidence of Fe_2O_3 and MnO enrichment near the paleosurface, suggesting the presence of some free oxygen in the atmosphere.

Pettijohn (1970) explained the lack of red sedimentary rocks in the lower 3 groups of the Huronian Supergroup by assuming that these groups represented marine sediments. This assumption may explain the lack of red colours but it falls short of explaining the presence of abundant detrital pyrite and uraninite minerals in lower Huronian quartz-pebble conglomerates. Indeed, these minerals would most probably be expected to survive deposition if deposited in a subaerial or fluvial environment exposed to an oxygen-poor atmosphere.

METAMORPHISM OF THE HURONIAN SUPERGROUP

During the Penokean Orogeny (1.7 to 1.9 Ga, Cannon 1973; 1.7 to 1.75 Ga, Hurst and Farhat 1978; 1.8 to 1.9 Ga, Medaris 1983), the rocks of the Huronian Supergroup were subjected to low-pressure, dynamothermal metamorphism to a maximum grade equivalent to the staurolite zone of the amphibolite facies. There is also evidence for another, more subtle, metamorphic event which preceded the emplacement of the Nipissing intrusive rocks (Card 1978a).

Subgreenschist- to lower greenschist-facies assemblages predominate in both the Sault Ste. Marie–Elliot Lake area of the Penokean Fold Belt and the Cobalt Embayment. The area south of the Murray Fault System, between Sudbury and Spragge, underwent a higher degree of deformation and is metamorphosed from middle greenschist to amphibolite facies (Figure 14.24; Card 1978b; Fleet et al. 1987).

On a regional scale, there is a general, positive correlation between metamorphic grade and the inferred, original thickness of supracrustal rocks. Areas of highest-grade metamorphism also coincide with major faults and intense folding; in detail, however, mapped isograds cross fold axes, and the style and intensity of deformation does not apparently correlate with the metamorphic grade. "Nodes" of higher-temperature metamorphic assemblages are present within low metamorphic grade areas (Card 1964, 1978b).

The abrupt change in metamorphic grade across the Murray Fault System is especially obvious in the Cutler–Spragge area where relatively undeformed, lower greenschist-facies, sericitic arenites situated north of the Murray Fault System, lie within a few hundred metres of retrograded, staurolite-bearing schists of the McKim Formation lying south of the fault system (Robertson 1970). Rocks of the Huronian Supergroup near the Creighton Pluton exhibit ptygmatic granitic mobilizate suggestive of incipient melting (Figure 14.25).

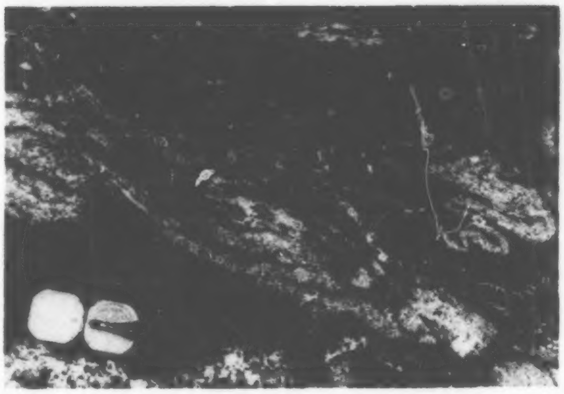


Figure 14.25. Ptygmatic granitic mobilizate in metasediment of the Huronian Supergroup near the Creighton Pluton, 3 km north of Meatbird Lake and 2.5 km east-northeast of Creighton Mine, in Snider Township (*from* Dressler 1984).

INTRUSIVE ROCKS SPATIALLY ASSOCIATED WITH THE HURONIAN SUPERGROUP

Several igneous intrusions are spatially associated with rocks of the Huronian Supergroup. These intrusions range from layered, gabbroic and granitic plutons, to diabase sills. Magmatism is probably related to the earliest Huronian volcanic events: 1) the Blezardian (Stockwell 1982) or pre–Penokean (>2.2 Ga) orogenic event; 2) the convergent tectonic events of the Penokean Orogeny; and 3) subsequent anorogenic, deep-crustal processes.

EAST BULL LAKE GABBRO-ANORTHOSITE INTRUSION

Layered intrusions of gabbro-anorthosite are found near, or at, the base of the Huronian Supergroup at several locations: from Falconbridge Township, southeast of the Sudbury Igneous Complex, to as far west as Gerow Township near the east end of the Quirke Lake Syncline.

Intrusions consisting of coarse-grained gabbro, anorthositic gabbro, gabbroic anorthosite and metamorphosed equivalents (Card 1976a; Robertson 1976) display locally well-developed, primary magmatic layers from a few centimetres to several tens of metres thick. Segregations, dikes and sheet-like bodies of granophyre are, in places, associated with the gabbro-anorthosites, and form a significant part of intrusions south of Agnew Lake (situated north of Espanola).

Major intrusions are at East Bull Lake, in Boon and Gerow townships (Born and James 1978; Kamineni et al. 1984); around Agnew Lake, in Shakespeare and Dunlop Townships (Card and Palonen 1976); and in the Massey area (Robertson 1976). Gabbroic rocks intersected in drill holes under Lake Huron in the Thessalon area, and an anorthositic-gabbro body on the shore of Lake Huron, in the Blind River area, possibly belong to this group of intrusions.

Krogh et al. (1984) reported an age of 2491±5 Ma for the gabbro-anorthosite intrusion in Shakespeare and Dunlop townships and a date of 2480^{+10}_{-5}Ma for the East Bull Lake Gabbro-Anorthosite Intrusion. This date is similar to the 2450^{+25}_{-10}Ma (zircon) age of the Copper Cliff Formation rhyolite (Krogh et al. 1984) and supports Card's (1978a) suggestion of a relationship between the gabbro-anorthosite bodies and Huronian volcanism. The Matachewan diabase dikes, so prominent in the shield north of the outcrop area of the Huronian Supergroup, have an age similar to the gabbro-anorthosite intrusions; as suggested by Fahrig (1987), the Matachewan dike swarm may be genetically related to the Huronian volanic rocks. However, Dressler (1984) found no evidence of intrusive relationships between gabbro-anorthosite bodies and the Huronian rocks southwest of the Sudbury Igneous Complex. Instead, he suggested a disconformable relationship as indicated by his observations of mafic dikes which were found to intrude the gabbroic rocks but which were not found to intrude the adjacent Huronian strata . This suggests that the gabbros are older than the Huronian rocks and is supported by Dressler's (1984) observation of Huronian conglomerate containing coarse-grained anorthosite clasts derived from one of the nearby gabbro-anorthosite bodies. Kamenini (Atomic Energy Corporation Limited, personal communication, 1990) also identified a disconformable relationship between the East Bull Lake Gabbro-Anorthosite Intrusion and overlying Huronian strata. If the upper surface of the gabbro-anorthosite intrusion is erosional, it is most likely equivalent to the sub–Matinenda Formation disconformity which lies, in part, on the Thessalon Formation.

PROTEROZOIC GRANITIC PLUTONS OF THE SUDBURY AREA

Three granitic plutons intrude Huronian volcanic rocks along the southeastern boundary of the Sudbury Igneous Complex. They are, from east to west, the Skead, Murray and Creighton plutons.

The Murray and Creighton granites have apophyses which extend into the surrounding volcanic rocks, and also have inclusions of the latter within the plutons themselves. The relationships of the Skead Granite with the Huronian supracrustal rocks is unclear, as no intrusive relationship with either the Huronian rocks or the Archean granites has been observed. There is field evidence that the Creighton and Murray plutons are older than the Sudbury Igneous Complex. Contact-metamorphic remobilization of the granites by the Sudbury Igneous Complex has resulted in rheomorphic dikes of the "older" Creighton and Murray plutons occurring within the younger Sudbury Igneous Complex. These dikes have previously been described by Burrows and Rickaby (1930) and Yates (1938).

Uranium-lead zircon ages of 2333 Ma for the Creighton Pluton (Frarey et al. 1982) and of 2388 Ma for the Murray Pluton (Krogh et al. 1984) show that they predate the 1850 Ma age (Krogh et al. 1984) assigned to the Sudbury Igneous Complex. These plutons may be related to the felsic volcanism of the 2450^{+25}_{-10} million-year-old (Krogh et al. 1984) Copper Cliff Formation.

NIPISSING INTRUSIVE ROCKS

Gabbroic rocks, first referred to as "Nipissing Diabase" by Miller (1913), are the most abundant and widespread igneous rocks intruding the Huronian Supergroup. Nipissing intrusive rocks form dikes, sills, undulating sheets up to several hundred metres thick and bodies which are interpreted as cone sheets (Lovell and Caine 1970; Card and Pattison 1973). The distribution of these tholeiitic intrusions is nearly uniform across the Penokean Fold Belt and the Cobalt Embayment. They are rarely recognized north of present limits of the Huronian Supergroup exposures. The areal extent of the Nipissing bodies is mainly the result of lateral movement of mafic magma along dikes and subhorizontal zones of weakness in the Huronian units (Fahrig 1987). There is evidence in the Quirke Lake Syncline for tectonic events prior to the intrusion of the Nipissing gabbro (Robertson 1962); this will be discussed in a subsequent section.

Corfu and Andrews (1986) reported a U-Pb baddeleyite age of 2219.4±4 Ma from wall-rock alteration related to intrusion of Nipissing gabbro sills from the Gowganda area. The baddeleyite is significantly older than the previously reported Rb-Sr ages of 2110±80 Ma (Van Schmus 1965) or 2116±27 Ma (Fairbairn et al. 1969), but the discrepancy is most likely the result of postemplacement disturbance of the Rb-Sr system. Buchan and Card (1985) suggested the presence of at least 2 periods of Nipissing intrusion as ascertained from paleomagnetic data. The paleopole of one period of Nipissing intrusion is similar to the paleopole of the Preissac dikes suggesting that the Nipissing dikes may be about the same age as the Preissac dikes (cf. Fahrig and West 1986; Osmani, this volume). Field relations suggest that the dikes postdate the sills.

Two-pyroxene gabbro appears to be the most common Nipissing intrusive-rock type. Other varieties of Nipissing intrusive rocks consist of olivine gabbro, hornblende gabbro, feldspathic pyroxenite, leucogabbro, granophyric gabbro and granophyre (Hriskevich 1968; Card and Pattison 1973).

In the Sudbury area, all except the central parts of some thick intrusions have undergone metamorphism from lower greenschist to lower almandine-amphibolite facies. In the metagabbros, pyroxene and calcic plagioclase have been replaced by amphibole, sodic plagioclase, epidote, talc, chlorite and quartz (Card and Pattison 1973).

A leucocratic, fine- to medium-grained, granophyric phase is common in the upper parts of many Nipissing bodies in the Cobalt Embayment. In the Sudbury area (Innes and Colvine 1979), and in the Iron Bridge area (Robertson 1963), steeply dipping, nonfoliated metagabbro bodies are dominant.

Irregular masses and dikes of mafic to intermediate pegmatite, consisting of amphibole and altered plagioclase crystals up to 10 cm long form a minor part of the Nipissing intrusions of some areas.

Bowen (1910) interpreted the granophyric phases of the Nipissing intrusions as the result of contamination. However, Card and Pattison (1973) and Hriskevich (1968) found compositional variation within Nipissing intrusions that is consistent with crystal fractionation.

No identified major tectonic event has been correlated with the Nipissing intrusive rocks. Fahrig (1987) proposed that the Nipissing intrusions belonged to the "Logan–Copper Cliff Node" and would represent 2 or more periods of basaltic magma emplacement along a failed arm perhaps related to the Animikie Basin or the main part of the Penokean orogen to the south.

Mineralization environments associated with the Nipissing intrusive rocks are described by Innes and Colvine (1984), Conrod (1988), Finn et al. (1982), and by Fyon et al. (this volume).

CUTLER PLUTON

The Cutler Pluton is located along the North Channel of Lake Huron about 35 km east of the town of Blind River. It is a foliated, composite, muscovite-biotite granitic body about 3 by 28 km in size. It intrudes both metamorphosed rocks of the McKim Formation and Nipissing intrusive rocks south of the Murray Fault System (Card 1978a; Robertson 1970) along the axis of a doubly plunging anticline (Cannon 1962; Robertson 1970).

The composite pluton consists of medium- to coarse-grained, foliated quartz monzonite, granodiorite and tonalite. It has the characteristics of an S-type granite (Condie 1989).

Discordant Rb-Sr and discordant K-Ar ages derived from micas and feldspars obtained from the granitic rocks and associated pegmatites indicate an age of about 1.7 Ga for the pluton (Wetherill et al. 1960). The Cutler Pluton was emplaced during the late phases of the Penokean Orogeny.

CROKER ISLAND AND MANITOULIN ISLAND COMPLEXES

A circular, composite, postorogenic pluton of granitic and syenitic to dioritic-gabbroic rocks intrudes quartz arenite of the Lorrain Formation on a group of islands—including Croker Island, Fox Island and the Benjamin Islands—in Lake Huron about 20 km south of the community of Massey. Collins (1925) referred to the intrusion as the "Eagle Island

Batholith". Card (1965b) described the intrusion and named it the "Croker Island Complex".

The pluton is about 10 km in diameter and coincides with one of several distinct aeromagnetic anomalies. Drilling on some of these anomalies on northern Manitoulin Island identified the presence of additional plutons, similar to the Croker Island Complex, covered by Paleozoic rocks (Van Schmus et al. 1975).

Van Schmus (1965) reported a Rb-Sr age for the Croker Island Complex of 1475±50 Ma. Van Schmus et al. (1975) analyzed drill core from a related intrusion covered by Paleozoic rocks on Manitoulin Island and reported a U-Pb age obtained for a quartz monzonite of 1500±20 Ma.

Although much of the Croker Island Complex is covered by water, Card (1965b) recognized concentric zoning of various rock types within the intrusion. Coarse-grained to porphyritic granitic rocks occur mainly near the margin of the pluton and as dikes in the core. Equigranular to porphyritic syenitic rocks predominate in the core of the complex and on adjacent islands. Dioritic and gabbroic rocks occur near the core and as dikes in the surrounding, granitic rocks (Card 1965b).

Card (1965b) stated that the emplacement of the pluton postdated the main tectonic and metamorphic events which affected the Huronian rocks of the area. The magma intruded in at least 3 successive pulses: an early gabbroic magma, a magma of intermediate composition and, finally, a granitic magma. The rocks range in composition from calc-alkalic gabbro to granite (Card 1965b).

PALEOENVIRONMENTAL SETTING OF THE HURONIAN SUPERGROUP

The paleoenvironmental setting of the Huronian Supergroup has been the subject of debate since the early years of the twentieth century, commencing when Coleman (1905a) proposed a glacial origin for the Gowganda Formation. With the development of plate-tectonic models, there have been attempts to view the origin of the Huronian Supergroup in terms of the processes applied to Phanerozoic crustal regimes. Tectonic proposals put forward by Young (1983) and Sims et al. (1981) are not radically different but will be described separately.

Sims et al. (1981) proposed that the Paleoproterozoic sedimentary basins of the Canadian Shield were the result of tectonic processes unique to the Paleoproterozoic. They suggested that a Paleoproterozoic basin developed along a zone of weakness between a Neoarchean (2750 to 2600 Ma) granite-greenstone complex and a Meso- to Paleoarchean (3500 to 3000 Ma) gneiss terrane to the south. These 2 crustal segments of differing tectonic stability were joined together in Archean time. Faulting and subsequent foundering along this zone of weakness—the Great Lakes Tectonic Zone—produced an intracratonic basin into which detritus derived from the Archean craton to the north was deposited.

Rocks of the Hough Lake and Quirke Lake groups are considered by Sims et al. (1981) to be the result of major regressive marine cycles caused by the tectonic instability of the southern block. The Ramsay Lake and Bruce formations' polymictic conglomerates are interpreted as deep-water, submarine debris flows initiated by the sudden increase in paleoslope brought on by syndepositional faulting. These coarse deposits were eventually covered by deep-water siltstone and mudstone, including turbidite (McKim and Pecors formations). A gradual return to shallow-water conditions led to the deposition of shallow-water marine arenites of the Mississagi and Serpent formations. Sims et al. (1981) accepted the glacial origin of the Gowganda Formation.

Young (1983) suggested that the Huronian Supergroup, along with the Paleoproterozoic supracrustal rocks of the Lake Superior area, formed part of an east-trending aulacogen which opened to an ocean in an area now occupied by the Grenville Province. It was also proposed that the Cobalt Group of the Huronian Supergroup is stratigraphically correlative with the Chocolay Group of the Marquette Range Supergroup, south of Lake Superior. Young (1983) stressed the importance of eastward thickening of the McKim, Pecors and Mississagi formations and concluded that this, together with the paleocurrent data, supported an aulacogen model for the Huronian Basin.

STRUCTURE

The supracrustal rocks of the Huronian Supergroup onlap towards the north over the Archean Superior Province. In the southeast, the Grenville Front Tectonic Zone (Easton, this volume) forms a tectonic boundary to the Huronian rocks. Lake Huron and areas underlain by Paleozoic sedimentary rocks lie to the south of most of the Huronian Supergroup. The Huronian Supergroup is not found south of the Manitoulin Island discontinuity (Van Schmus et al. 1975). Huronian rocks between Sault Ste. Marie and Sudbury were subjected to folding and faulting both after intrusion of the Nipissing dikes (Card 1976a), in the so-called Blezardian Orogeny (*circa* 2.2 Ga, Stockwell 1982), and during the Penokean Orogeny. Distinction of 2 orogenic pulses based on geochronologic or structural evidence has not yet been made possible, in our view. Therefore, the following structural discussion does not attribute structures to specific Proterozoic deformations.

In the Penokean Fold Belt between Sault Ste. Marie and Sudbury, Huronian rocks display easterly trending, regional-scale folds and several faults or fault zones. Regional folding also appears to have affected Huronian sequences below the Gowganda Formation both in the Elliot Lake area (Collins 1925; Robertson 1961, 1962, 1968) and in the southern Cobalt Embayment (Card et al. 1973; Dressler 1986). Very limited information is available on the structure of this area; no description is offered here. Northeast of Sudbury, in the Cobalt Embayment, the tectonic style of pre–Gowganda Formation rocks is different from that of the Cobalt Group. Here in the Cobalt Embayment, rocks of the Cobalt Group contain large open folds and are cut by regional faults. The rocks below the Gowganda Formation show a more complex structural style. The present summary deals with the Huronian Supergroup between Sault Ste. Marie and Sudbury, described herein as the Penokean Fold Belt

In the McGregor Bay–Agnew Lake section across the Penokean Fold Belt (*see* Figure 14.26b), south-dipping synclines and anticlines, strike-parallel faults and discordant transcurrent faults reveal a complex structural picture (Card 1978a) suggestive of north-northwest-directed compression. Folds are commonly open, and the axial planes are upright to somewhat northward overturned.

In the northern section of this part of the Penokean Fold Belt, *en échelon*, doubly plunging synclines and anticlines form major synclinoria and anticlinoria and have axial surfaces that dip vertically to 80° (Card 1978a). In the southern part of this section near the Grenville Front Tectonic Zone, Card (1978a) described an irregular fold pattern suggestive of refolding due to interference between east-trending structures and northeast-trending structures parallel to the Grenville Front.

Folds in the McGregor Bay–Agnew Lake area of the Penokean Fold Belt are of several ages and orientations. Those spatially associated with of the Murray Fault System are prominent. They trend east to northeast and, according to Card (1978a) and Robertson (1970), their axial surfaces probably dip steeply southward. Repetitions and omissions of rock units indicate major displacements on faults. Robertson (1970) and Card (1978a) described reverse, dip-slip (south-side up) as well as dextral, strike-slip movement on these faults and stated that the Murray Fault System was active over at least 1 billion years. Dip-slip movements of approximately 10 to 15 km on the Murray Fault System are indicated by metamorphic data. Zolnai et al. (1984) estimated that south of the Murray Fault System, Huronian rocks represent midcrustal levels (15 to 20 km depth), whereas north of the fault system, burial was to about 5 km depth. Minor folds and shear-zone fabrics in mylonitized sedimentary and volcanic rocks within the Murray Fault System indicate a dextral-sense of ductile shear (Lorek 1987).

Northeast- and northwest-trending faults probably belong to a conjugate set of normal faults; they have produced appreciable offsets within the Huronian rocks "of a few hundred to a few thousand feet" (Card 1978a).

Zolnai et al. (1984)(Figure 14.26) made observations along 2 regional transects conducted across the Penokean Fold Belt north of Lake Huron. One of these transects, in the McGregor Bay–Agnew Lake area, was also investigated by Card (1978a), the other transect was in the Elliot Lake area. No décollement was observed by Zolnai et al. (1984) between the rocks of the Huronian Supergroup and the Archean basement rocks of the Superior Province at either the northernmost boundary of the Huronian Supergroup or around the Quirke Lake Syncline. This suggests that basement and cover rocks were folded together.

The Flack Lake Fault (*see* Figure 14.26), based upon tectonic (Zolnai et al. 1984) and paleoenvironmental (Roscoe 1957a, 1957b; Robertson 1971) considerations, follows the locus of an older normal fault and is interpreted to be a reverse, listric thrust fault. Abrupt southward thickening of rock units occurs across several faults between the Flack Lake Fault and the Murray Fault System. A distinct paleoenvironmental facies change is marked by the Murray Fault System, indicating that it was active during and after deposition of the Huronian rocks. North of the fault system, the rocks are interpreted as fluvial; those to the south are deep-water turbidites (Card et al. 1972). Zolnai et al. (1984) interpreted the Murray Fault System to be a fault-controlled hinge zone "south of which rocks that accumulated to a greater thickness were subsequently tectonically buried to middle crustal depth before being thrust up and over adjacent rocks on the north that were never very thick or deeply buried and had remained unmetamorphosed".

West of the Elliot Lake and the McGregor Bay–Agnew Lake sections of the Huronian Supergroup (sections A-A′ and B-B′, respectively, on Figure 14.26a and 14.26b) towards Sault Ste. Marie, the structural style (Frarey 1977) is very similar to the one described above; however, no detailed structural work has been done in this area.

Recent studies by Shanks (*see* Dressler et al., this volume), on the Sudbury Structure, identified a large, reverse (south over north), ductile-deformation zone extending westward from the Sudbury Structure. Shanks (1990) speculated that the zone may be a splay of the Murray Fault System.

Several large-scale interpretations of the Penokean Fold Belt and its tectonic setting within central North America have been attempted. Sims et al. (1980) considered the fold belt as part of the Great Lakes Tectonic Zone that affects both Proterozoic and Archean rocks westward from Sudbury to west of Minnesota. This tectonic zone separates the Wawa Subprovince, on the north, from the Marquette greenstone belt and the Minnesota River Valley gneisses on the south. Tectonism along the Great Lakes Tectonic Zone culminated during the Paleoproterozoic Era (1850 to 1950 Ma) after structural basin development.

Hoffman (1989) presented a large-scale tectonic interpretation of the area under consideration here as a continental-margin sedimentary sequence shortened by north-directed thrusting during the Penokean Orogeny, caused by the accretion of the Pembine–Wausau Terrane, one of several Paleoproterozoic accreted terranes flanking an Archean protocraton.

In the following account, we propose an integrated tectonic and depositional model for the Huronian Supergroup (Figures 14.27a to 14.27e) which draws heavily on the work of Zolnai et al. (1984), Barovich et al. (1989), Young (1983) and Roscoe (1969). Our reconstruction of these complex geological events includes areas of speculation which reveal our prejudice and geological experience.

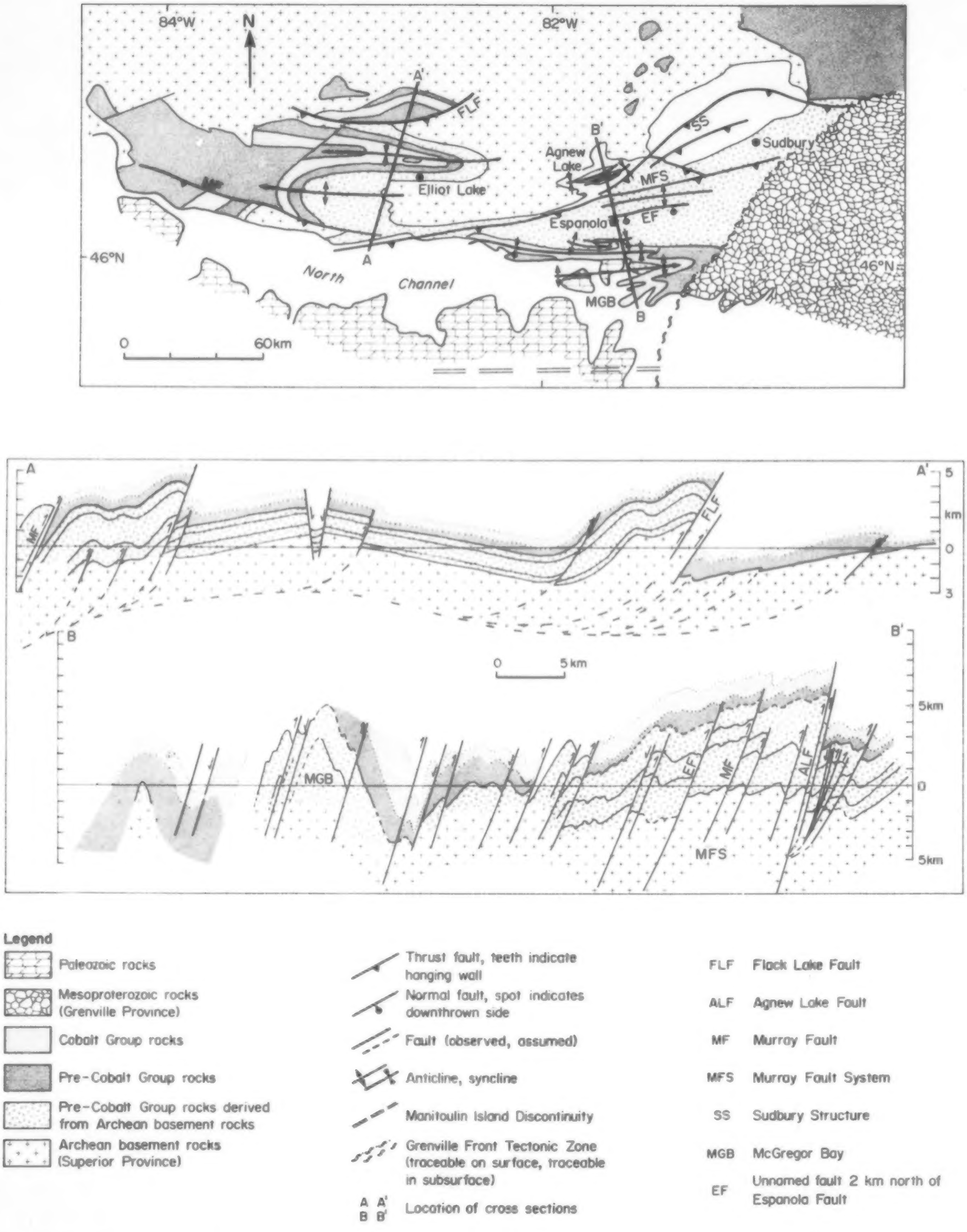

84°W
N
82°W
A'
FLF
SS
B'
Sudbury
Agnew Lake
Elliot Lake
MF
MFS
Espanola
EF
46°N
North Channel
A
MGB
B
46°N
0
60km
A
A'
5
km
0
3
MF
FLF
B
B'
0
5 km
5km
0
5km
MGB
EF
MF
ALF
MFS
Legend
Paleozoic rocks
Mesoproterozoic rocks (Grenville Province)
Cobalt Group rocks
Pre-Cobalt Group rocks
Pre-Cobalt Group rocks derived from Archean basement rocks
Archean basement rocks (Superior Province)
Thrust fault, teeth indicate hanging wall
Normal fault, spot indicates downthrown side
Fault (observed, assumed)
Anticline, syncline
Manitoulin Island Discontinuity
Grenville Front Tectonic Zone (traceable on surface, traceable in subsurface)
A A' B B' Location of cross sections
FLF Flack Lake Fault
ALF Agnew Lake Fault
MF Murray Fault
MFS Murray Fault System
SS Sudbury Structure
MGB McGregor Bay
EF Unnamed fault 2 km north of Espanola Fault

Figure 14.26. a) Geological sketch map, and b) regional north-south cross section of the Penokean Fold Belt, north of Lake Huron (*from* Zolnai et al. 1984).

A Tectonic Model for the Huronian Supergroup

DIVERGENT TECTONIC REGIME

Passive Rift Stage

Livingstone Creek Formation: earlier than 2.48 Ga, later than ~2.65 Ga, *see* Figure 14.27a.

The abundance of coarse-grained, polymictic conglomerate, rapid changes in stratigraphic thickness, and a local source for the Livingstone Creek Formation all suggest deposition in fault-bounded basins, some of which had north-facing paleoslopes. There is no evidence of Huronian volcanism either prior to, or during, the deposition of the Livingstone Creek Formation.

It has been argued earlier in this review that a period of erosion followed the deposition of the Livingstone Creek Formation. This erosional interval could permit the Livingstone Creek Formation to have had a much greater areal distribution than is presently inferred.

We speculate that the Livingstone Creek Formation was deposited in a passive rift configured by a pre-existing zone of weakness in the Archean crust. There is no volcanism directly associated with the Livingstone Creek Formation, so we speculate that the Huronian Basin may have been initiated as a passive rift—that is, a rift without associated volcanic rocks—located along an older structure (Condie 1989). The Great Lakes Tectonic Zone is such an older structure (Sims et al. 1980). The parallel alignment of Archean structural trends and the Huronian Supergroup, in combination with Jolly's (1987) conclusion that the geochemical signature of the Thessalon Formation's volcanic rocks is consistent with mantle metasomatism associated with Archean subduction, together suggest that Archean events may have influenced the orientation and character of Huronian Basin development.

Active Rift Stage

Volcanic Rocks of the Huronian Supergroup: 2490 to 2480 Ga, *see* Figure 14.27b.

The limited areal extent of the Livingstone Creek Formation suggests that doming may have preceded active rifting. Such a condition would have resulted in the erosion of much of the Livingstone Creek Formation in the present Quirke Lake Syncline and possibly to the east of it.

The development of the passive rift into an active rift is indicated by the initiation of volcanism. Volcanic activity began with central-vent eruptions, as represented by the lower Thessalon Formation.

The upper tholeiitic basalt part of the Thessalon Formation, along with the Elsie Mountain and Stobie formations of the Sudbury area, suggest major eruption of flood basalt during the main period of active rifting. The Thessalon Formation was partly eroded from the rift scarps—and perhaps from an area of unknown extent east of the Quirke Lake Syncline—as a result of regional doming or local uplift in the rift zone. The rift association of the Huronian volcanic rocks is based in part on the transitional alkalic character of the volcanic rocks in the Sudbury area.

The East Bull Lake suite of gabbro-anorthosite intrusions appears to be, as Card (1978a) has suggested, closely related to Huronian volcanism. The suite probably represents high-level magma chambers.

Fahrig (1987) suggested that the Matachewan dike swarm represents a failed arm of the Huronian rift.

Late Rift or Early Passive-Margin Stage

Salmay Lake, Elsie Mountain, Stobie, Matinenda and McKim formations: 2450 Ma, *see* Figure 14.27c.

The Huronian rocks themselves provide little direct evidence of rifting of the Archean continent and of the creation of new oceanic crust. Valid arguments may be made for an intracontinental setting for the Huronian Basin. However, we believe that many of the lithological and structural elements of the Huronian Supergroup may be most conveniently interpreted in terms of divergent and convergent tectonic settings of a plate-tectonic regime.

The formation of both new oceanic crust to the south of the present Huronian outcrops and active "drifting" resulted in the cooling of the newly formed lithospheric mantle and concomitant collapse of the weakened and attenuated continental margin. The sands and gravels of the Matinenda Formation were deposited by streams flowing to the new continental margin over the partly eroded volcanic rocks of the Thessalon Formation. The marine sediments of the McKim Formation indicate that the floor of the rift had fallen below sea level.

Passive-Margin Stage

Hough Lake, Quirke Lake and Cobalt groups: 2.45 to 2.22 Ga, *see* Figure 14.27d.

The majority of the rocks of the Huronian Supergroup belong to this stage. From this point onward, sediment transport and deposition was largely determined by climatic changes and tectonic activity north of the Penokean Fold Belt. The listric normal faults of the rift stages were reactivated in response to the sediment load.

The cycles of conglomerate, mudstone-siltstone and arenite, which define the upper 3 Huronian groups, are interpreted as responses to periodic episodes of continental glaciation. In this scenario, glaciers carry unsorted detritus to a continental margin—the margin is undergoing isostatic adjustment to the ice load—resulting in a marine transgression with deposition of marine sediments over the glacial till. Melting of the ice sheet, coupled with tectonic activity in the source area, could lead to widespread deposition of fluvial and shallow-marine sand over the marine muds and silts (Roscoe 1969; Casshyap 1969). The continuous distribution of the relatively thin, singular conglomerate units (Ramsay Lake and Bruce formations), and the presence of dropstones in the Pecors and Gowganda formations are more readily interpreted in terms of continental glaciation.

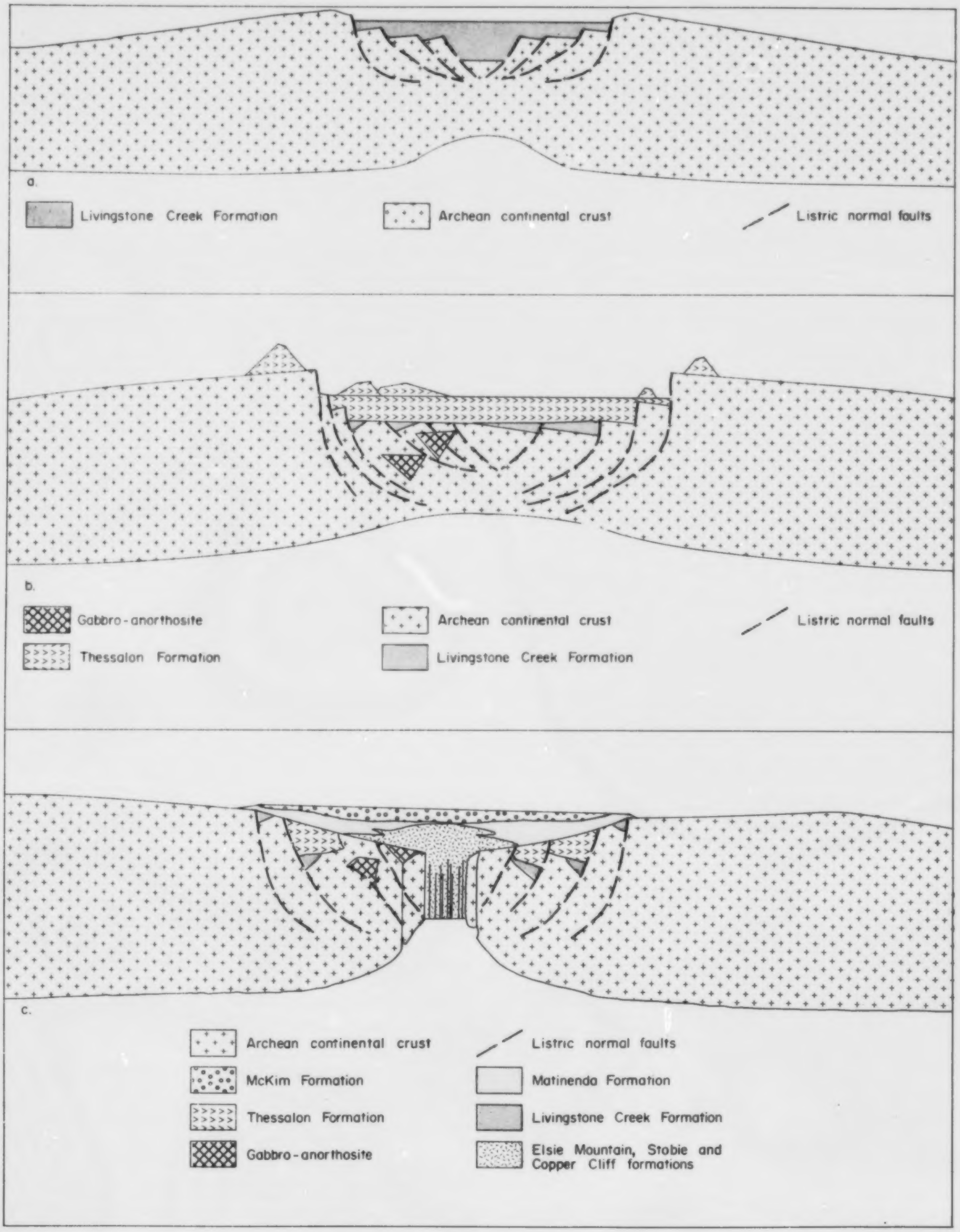

a.
Livingstone Creek Formation
Archean continental crust
Listric normal faults
b.
Gabbro-anorthosite
Thessalon Formation
Archean continental crust
Livingstone Creek Formation
Listric normal faults
c.
Archean continental crust
McKim Formation
Thessalon Formation
Gabbro-anorthosite
Listric normal faults
Matinenda Formation
Livingstone Creek Formation
Elsie Mountain, Stobie and Copper Cliff formations

Figure 14.27. Simplified schematic diagrams showing successive stages in the proposed paleotectonic model for the development of the Huronian Supergroup and its subsequent deformation and metamorphism. **a)** Initial passive rift stage: sediments of the Livingstone Creek Formation are deposited in graben(s). **b)** Main rift development stage accompanied by eruption of Huronian volcanic rocks. **c)** Late rift or early breakup stage: eruption of Sudbury area volcanic rocks (Eastern volcanic rocks) and deposition of sediments of Matinenda Formation; beginning of deposition of McKim Formation sediments.

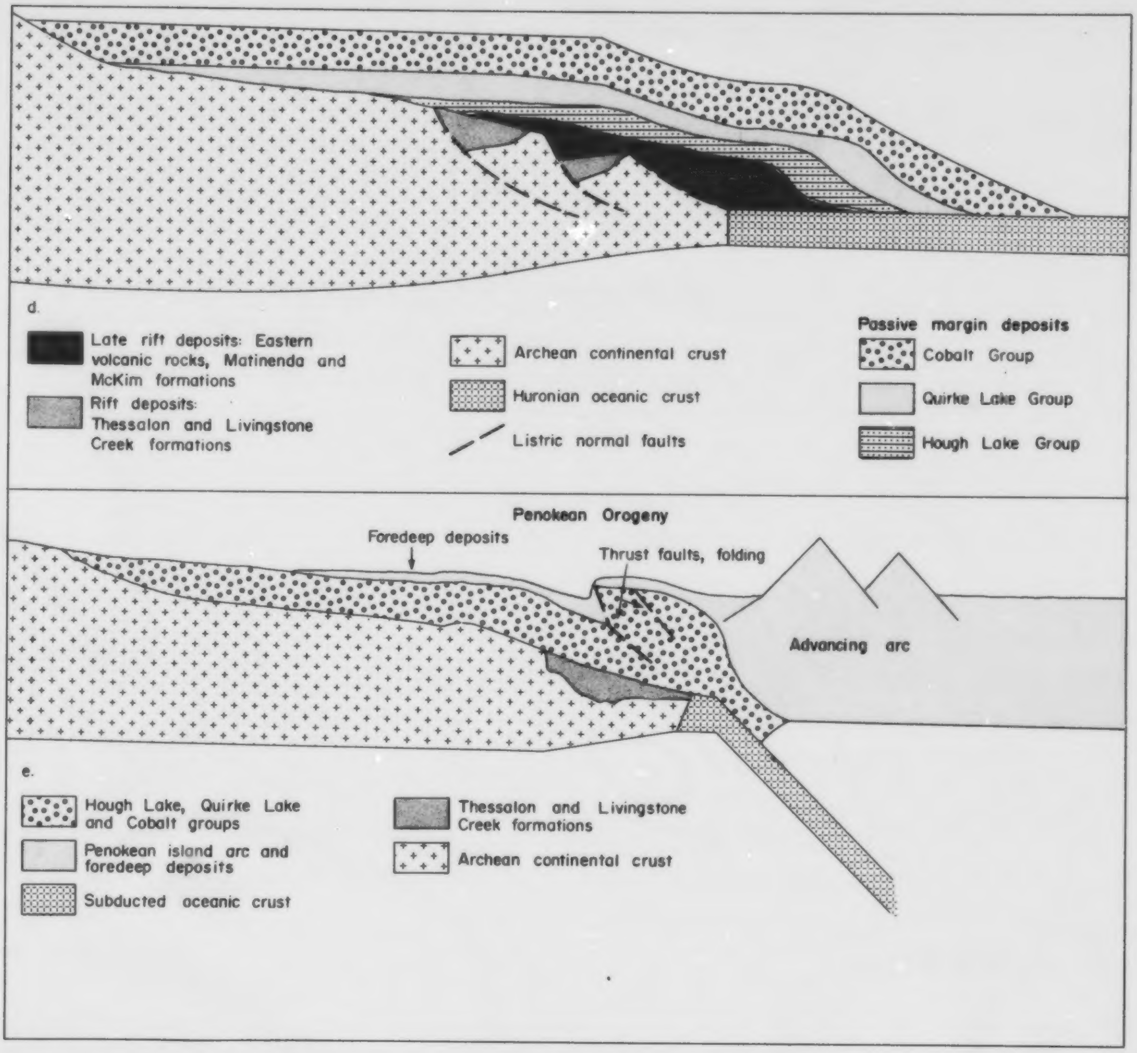

d.
Late rift deposits: Eastern volcanic rocks, Matinenda and McKim formations
Rift deposits: Thessalon and Livingstone Creek formations
Archean continental crust
Huronian oceanic crust
Listric normal faults
Passive margin deposits
Cobalt Group
Quirke Lake Group
Hough Lake Group
Penokean Orogeny
Foredeep deposits
Thrust faults, folding
Advancing arc
e.
Hough Lake, Quirke Lake and Cobalt groups
Penokean island arc and foredeep deposits
Subducted oceanic crust
Thessalon and Livingstone Creek formations
Archean continental crust

Figure 14.27. Simplified schematic diagrams showing successive stages in the proposed paleotectonic model for the development of the Huronian Supergroup and its subsequent deformation and metamorphism. **d)** Passive-margin stage: 3 successive periods of continental glaciation deposit sediments of the Hough Lake, Quirke Lake and Cobalt groups on an attenuated continental margin. **e)** Collision of continental block with an island arc and deposition of foredeep sediments of the Chelmsford Formation.

The proliferation of oxygen-producing phytoplankton during the period of Huronian Supergroup deposition led to the stability of ferric iron in rocks younger than the Quirke Lake Group (Frarey and Roscoe 1970).

The area from which the Huronian Supergroup's terrigenous sediments were derived lies in the direction of the Kapuskasing Structural Zone (Williams et al., this volume), but it is not clear if uplift along this structural zone coincided temporally with Huronian sedimentation. The uplift may have contributed to the present distribution of Huronian rocks; either by erosion and/or by control on sedimentation. The angular unconformity below the Cobalt Group in the northern part of the Huronian Basin area, in contrast with its being a minor disconformity near Lake Huron, points to a tectonic disturbance well north of the basin. The end of this fourth stage is possibly marked by the deformation which precedes intrusion of the 2.2 billion-year-old Nipissing diabase.

CONVERGENT TECTONIC REGIME

Collision Stage

Penokean Orogeny: 1.8±0.1 Ga, *see* Figure 14.27e.

The Huronian Supergroup has been subjected to a pre–Nipissing diabase tectonic event which produced folds that were eventually cut by the diabase (Card 1976a; Stockwell 1982). The causes for this event and its precise timing remain obscure. The region was subsequently subjected to a second orogenic event—the Penokean Orogeny (cf., Stockwell 1982).

Zolnai et al. (1984) proposed a plate-tectonic model for the observed structural features of the Huronian Supergroup rocks. They suggested that a block of Archean continental crust, advancing from the south, collided with, overrode and then depressed the southern part of the Huronian passive margin and rift sequences to midcrustal levels south of the Murray Fault System. Deep burial of supracrustal rocks as a consequence of tectonic thickening by folding, and due to an overriding allochthonous Archean terrane, led to Penokean ductile deformation and high-grade dynamothermal metamorphism of the Huronian rocks south of the Murray Fault System. Rocks north of the Murray Fault System were never deeply buried. Transportation of the deformed wedge of Huronian rocks over the Superior Province was progressive in a northward direction, producing listric thrust faults, such as the Flack Lake Fault, by reactivation of older faults. Ductile features south of the Murray Fault System were rotated into their present vertical attitudes by continued crustal shortening (Zolnai et al. 1984).

Zolnai et al. (1984) suggested that their inferred southern Archean block now lies beneath the Paleozoic sedimentary cover south of Manitoulin Island.

More recently, Barovich et al. (1989) have provided data on Nd isotope systematics of rocks from the states of Michigan and Wisconsin that support the proposal of Cambray (1978) and Larue (1983). Evidence by these workers suggests that the Penokean Orogeny of the Great Lakes region represents a collision between a Proterozoic island arc and the Superior protocraton. They concluded that the upper Michigamme Formation of the Marquette Range Supergroup has a Nd isotope signature consistent with its origin as a foredeep deposit derived from an island arc advancing from the south. The Penokean Orogeny was produced by the eventual collision of this arc with the Archean continent and early Proterozoic passive margin deposits (Chocolay Group).

The timing and common deformational and metamorphic features of the area south of Lake Superior and the Penokean Fold Belt suggests that the model of Zolnai et al. (1984) should be modified by the substitution of a Proterozoic island arc for the southern Archean continent.

The series of events listed above is comparable to actual plate-tectonic processes. There is, in fact, some doubt that major orogenies can occur without collision (Condie 1989). Some rocks which lie within the Penokean Fold Belt have ages and petrotectonic assemblages consistent with a collision orogen. For example, the syntectonic muscovite-bearing granite of the Cutler batholith displays many of the features of collision-related S-type granites (Condie 1989).

ECONOMIC GEOLOGY

Introduction

Many uranium, base metal and precious metal occurrences are associated with various rock types of the Huronian Supergroup or with the intrusive rocks spatially associated with the supergroup. Most of the deposits are small, but several are significant, especially those at Cobalt, Gowganda and Bruce Mines. The Huronian sequence is well known for its uranium deposits. The uranium deposits are described in some detail below. Silica is extracted from open pits located south of Espanola. For an account of other deposits within, or spatially associated with, the supergroup, the reader is referred to Fyon (this volume). Colvine (1981) and Long and Colvine (1985) discussed the gold potential of several Huronian formations.

Uranium

Uranium and thorium-uranium deposits are found in pyritiferous quartz-pebble conglomerate at a number of localities in the Blind River–Elliot Lake and Agnew Lake areas, as well as throughout the Penokean Fold Belt (Robertson and Gould 1982, 1983). Similar mineralization occurs in the Black Hills of South Dakota, in the Medicine Bow and Sierra Madre mountains of Wyoming (Houston and Karlstrom 1979; Roscoe 1981), and in the Witwatersrand Basin in South Africa (Pretorius 1981). In Canada, deposits at Sakami Lake and the Otish Mountains in Quebec, and Montgomery Lake in the Northwest Territories (Robertson, 1974; Roscoe 1981), also bear similarities in style and setting to the Blind River deposits. Studies on the Witwatersrand, Blind River and Jacobina (Brazil) deposits have formed the basis for definition of the uranium±gold-bearing, pyritic, quartz-pebble-conglomerate class of ore deposits (Houston and Karlstrom 1979; Armstrong 1981; Button and Adams 1981; Pretorius 1981).

BLIND RIVER–ELLIOT LAKE AREA

The uranium deposits of the Blind River–Elliot Lake area (Figure 14.28) are restricted to the basal parts of the Matinenda Formation. Host rocks are pyritic quartz-pebble conglomerate interbedded with conglomeratic to arkosic quartz arenite. The location and thickness variation of the sedimentary rocks are controlled by basement topography. In the ore zones, arkoses are thicker bedded and coarser grained than in unmineralized rocks; they assume a yellow-green colour upon weathering and commonly show scour-and-fill structures, cross-bedding and fining-upward cycles.

Locally, as seen in the mines on the northern limb of the Quirke Lake Syncline, there is an unconformity at the base of the conglomerate of the Ramsay Lake Formation; there, unconsolidated material from the Matinenda Formation has been incorporated into the Ramsay Lake Formation conglomerate. Consequently, this part of the Ramsay Lake Formation contains anomalous, but minor amounts of uranium (Robertson 1968, 1976).

The sedimentological environments of the conglomerate and arenite of the Matinenda Formation have been described by Fralick and Miall (1989). Studies by Theis (1979) have addressed lateral variation of clast size (at New Quirke and Denison mines on the northern limb of the Quirke Lake Syncline) within selected "ore reefs" (a mining term for mineable beds of uraniferous quartz-pebble conglomerate), the nature and distribution of the ore minerals and the distribution of elements of economic interest. Robertson (1968), Ruzicka and Steacy (1976) and Ruzicka (1981) have addressed the distribution of kerogen within the ore-bearing units. The kerogen, recently described by Willingham et al. (1985), was originally described as thucholite, possibly representing algal mats trapped at the top of fining-upwards units, that contained disseminated globules of organic material derived from these mats. The paleosols underlying the Huronian sequence are being studied to establish the general state of oxidation at the time of ore deposition.

The largest ore deposits of the Elliot Lake area are located in the Quirke Lake Syncline, but the discovery zone of the Pronto Mine lies some 20 km to the south of Elliot Lake. A cross section of the New Quirke Mine is shown in Figure 14.29. The ore zones, which are stratiform, strike northwest and are controlled by basement topography. The Quirke zone is the largest in the Elliot Lake area (9600 m long) and contains several mines. The Nordic Zone is 5800 m long. The Pronto deposit and the unworked zones are smaller (Robertson 1968, 1976, 1986). Total production, average grades, and resources are described below.

The Quirke and Nordic zones of the Elliot Lake mining camp, both containing more than one "ore reef", will be here selected as cases for discussion, as they are typical examples of the mineralization style occurring within this mining camp.

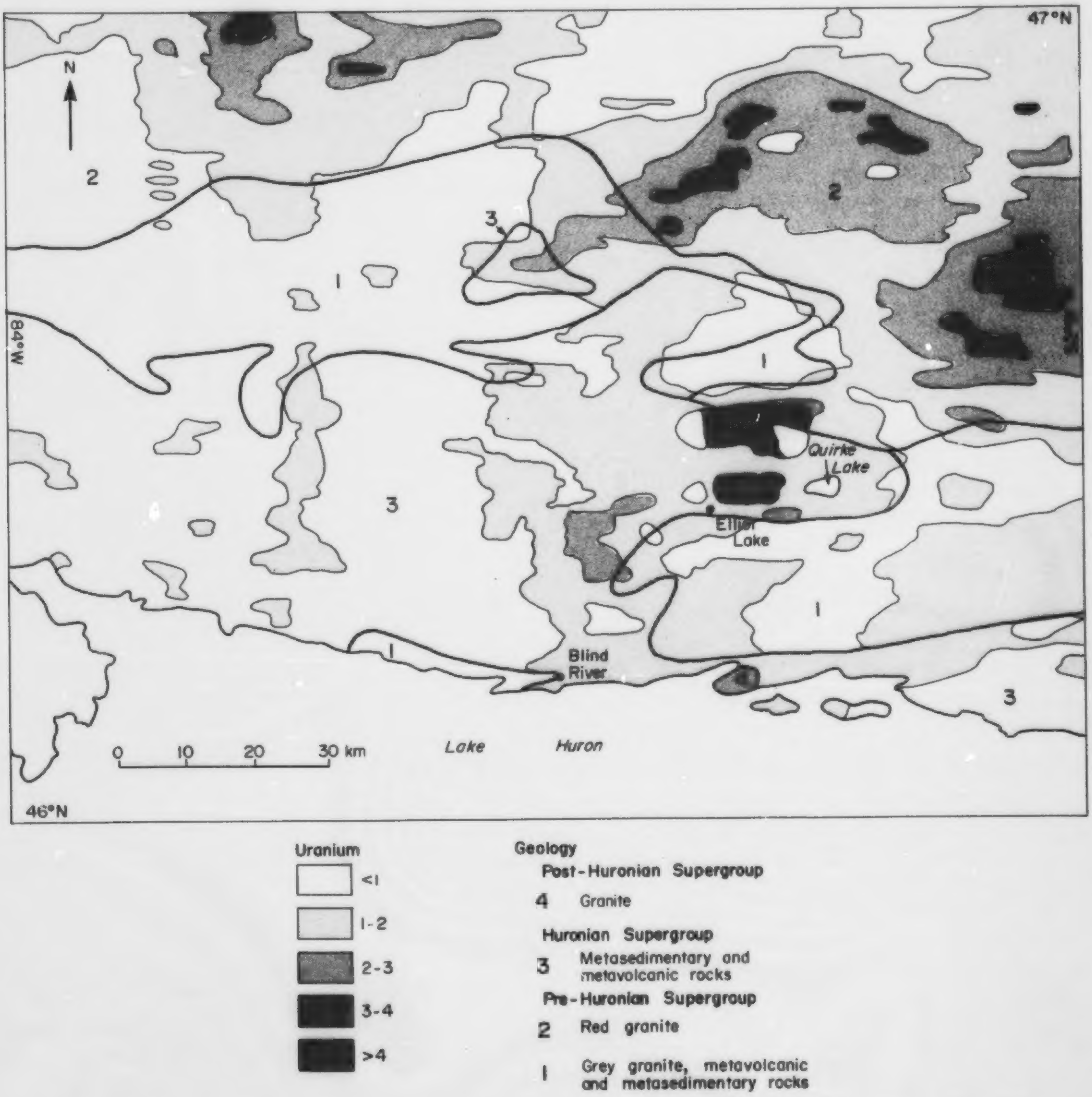

47°N
N
84°W
46°N
1
2
3
4
Quirke Lake
Elliot Lake
Blind River
Lake Huron
0
10
20
30 km
Uranium
<1
1-2
2-3
3-4
>4
Geology
Post-Huronian Supergroup
4 Granite
Huronian Supergroup
3 Metasedimentary and metavolcanic rocks
Pre-Huronian Supergroup
2 Red granite
1 Grey granite, metavolcanic and metasedimentary rocks

Figure 14.28. Uranium distribution as determined from airborne gamma-ray spectrometry surveys undertaken in the Blind River area. The two areas with anomalies of >4 ppm are areas in which mine tailings are stored (*modified from* Richardson et al. 1975).

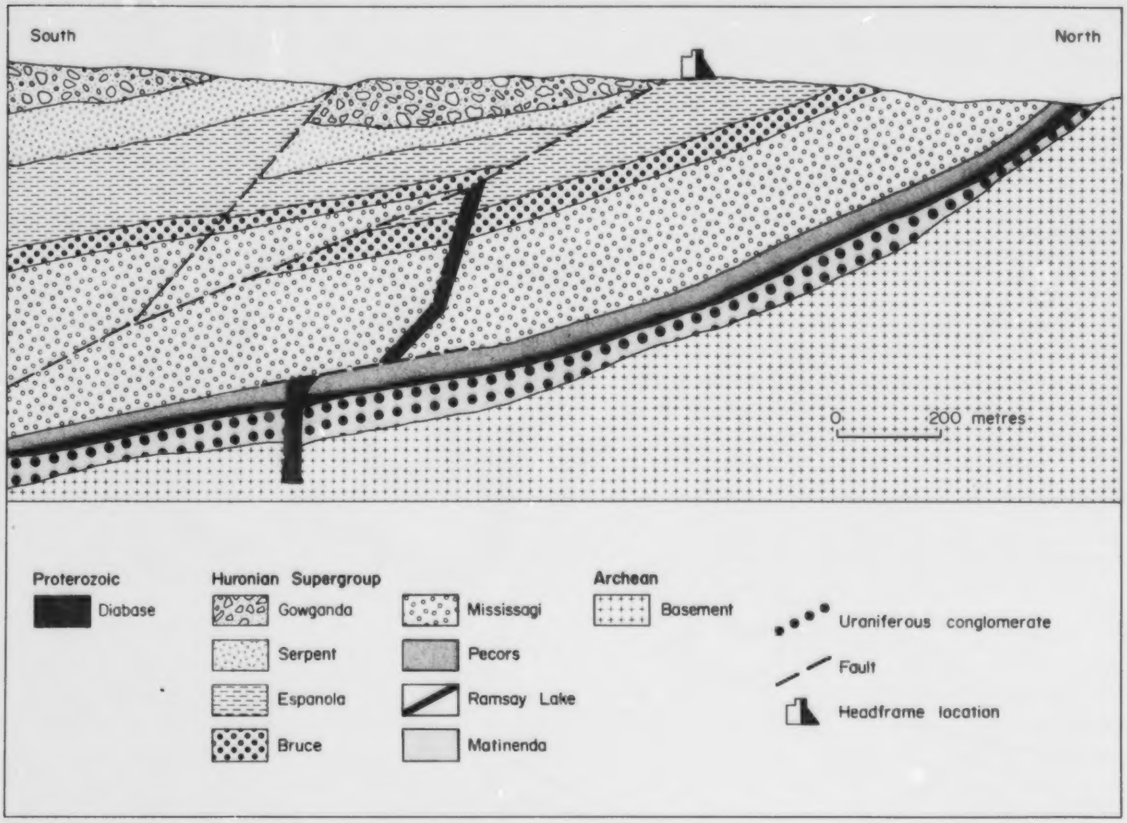


Figure 14.29. Cross section, New Quirke Mine, Elliot Lake area (*modified from* mine plans supplied by Rio Algom Ltd.).

At Quirke No.1 Mine, the main "ore reef" is approximately 30 m above basement and is approximately 3.5 m thick. Toward the east, the ore zone is truncated by an unconformity at the base of the Ramsay Lake conglomerate. At Quirke No.2, the richest ore-grade development is in the Denison Reef, some 30 m below the Quirke Reef. The Denison Reef normally comprises 2 conglomerate zones, each 1.8 to 3.6 m thick, separated by barren arkose 0.6 to 2.4 m thick. Other conglomerate beds, 0.6 to 3 m thick, separated by quartz arenite beds which are 3.6 to 6 m thick, are known on both the Quirke and Denison mine properties.

At the Nordic Mine, the main ore bed comprises either conglomerate or conglomerate with quartz arenite over a width of 3 m with a grade of 2.5 pounds U_3O_8 per ton (1.25 kg/t U_3O_8). Locally another "ore reef" lower in the sequence—the Lacnor Reef—was mined where grades attained 2.0 pounds U_3O_8 per ton (1 kg/t U_3O_8). In the eastern part of the Nordic Mine, a third "ore reef"—the Pardee Reef—attained 2.3 pounds U_3O_8 per ton (1.15 kg/t U_3O_8) over 1.5 m. This "ore reef" extends eastward over a basement ridge into the Pardee and Pecors mineralized zones. These "ore reefs" extend to the Stanleigh Mine where original operations were largely carried out over the Lacnor and Nordic reefs (*see* also Barnes 1986).

"Ore reefs" at Stanleigh Mine are named the lower, middle and upper reefs. As with the Quirke zone, other reefs of lower grade (which locally may average 1 pound U_3O_8 per 5 feet), or of minor extent, are known.

AGNEW LAKE AREA

Pyrite- and uranium-bearing quartz-pebble conglomerate beds in the basal Huronian rocks also occur at Agnew Lake (Figure 14.30). At the Agnew Lake deposit (Figure 14.31; Wilton 1986), the ore is composed of oligomictic, gritty-matrix, conglomerate interbedded with arkose in a metamorphosed, folded and deformed quartz arenite-argillite-conglomerate sequence which is probably equivalent to the Matinenda Formation of the Elliot Lake area. Pebbles at the Agnew Lake deposit are smaller and sparser, and pyrite is less conspicuous. The ratio of Th:U is generally higher than at Elliot Lake, reflecting the presence of substantial uranothorite and monazite. Post–Huronian deformation has been considerably greater, resulting in steeper dips, more pronounced cleavage, and elongation of pebbles. Permeability along the cleavage aided the extraction of uranium by leaching from broken ore.

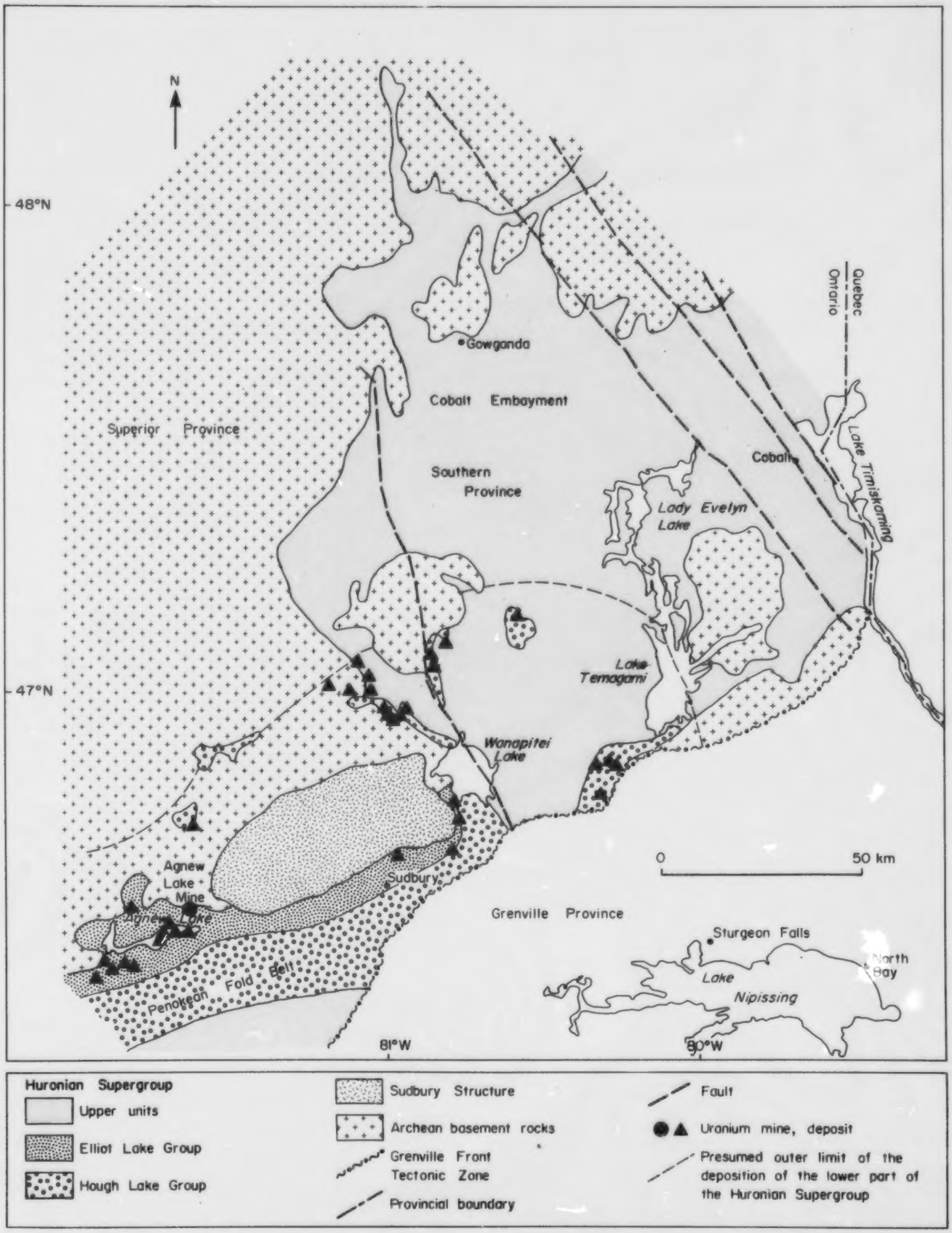

N
48°N
47°N
Gowganda
Cobalt Embayment
Superior Province
Southern Province
Cobalt
Lady Evelyn Lake
Lake Timiskaming
Quebec
Ontario
Lake Temagami
Wanapitei Lake
Agnew Lake Mine
Agnew Lake
Sudbury
Penokean Fold Belt
Grenville Province
Sturgeon Falls
North Bay
Lake Nipissing
0
50 km
81°W
80°W
Huronian Supergroup
Upper units
Elliot Lake Group
Hough Lake Group
Sudbury Structure
Archean basement rocks
Grenville Front Tectonic Zone
Provincial boundary
Fault
Uranium mine, deposit
Presumed outer limit of the deposition of the lower part of the Huronian Supergroup

Figure 14.30. Uranium occurrences in the Cobalt Embayment and in the Agnew Lake area of the Penokean Fold Belt.

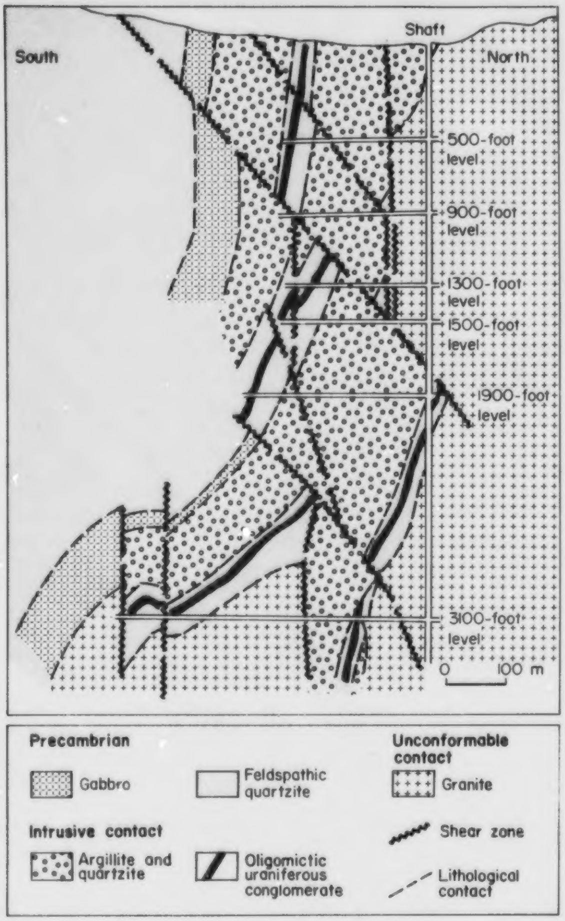


Figure 14.31. Cross section, Agnew Lake Mines Limited (*modified from* mine plans supplied by Agnew Lake Mines Limited).

COBALT EMBAYMENT

Lower Huronian Supergroup rocks are exposed in the southern part of the Cobalt Embayment. In the lowermost units, there are anomalous concentrations of uranium, and at some localities there is gold in conglomerate and uranium in argillaceous quartz arenite (*see* Figure 14.30; Meyn 1973; Meyn and Matthews 1980; Robertson 1981; Dressler 1986). Precise stratigraphic correlation of the lowermost stratigraphic units with similar units in the Blind River area is unclear, but rocks equivalent to the Mississagi and younger formations are present. Meyn and Matthews (1980) reported the uraniferous minerals as brannerite, uraninite and coffinite. They suggested a braided-stream environment for the deposition of the combined sediment and uranium but stated that some of the uranium could have been transported in solution.

LITHOLOGY AND MINERALOGY OF URANIUM DEPOSITS

The typical ore-bearing conglomerates of the Matinenda Formation consist of well-rounded, well-sorted quartz pebbles or cobbles set in a matrix of quartz, feldspar, and sericite and having a pyrite content of from 6 to 10%. Monazite and zircon are characteristic heavy minerals. Brannerite and uraninite are found in the matrix. Thucholite (kerogen) is found as disseminations and locally as thin beds and may occur in fissures in the ore beds. At Elliot Lake, the ore minerals are brannerite, uraninite, and monazite, with only traces of thorite and uranothorite, but the two latter minerals are characteristic of the Agnew Lake ores (Wilton 1986).

Patchett (1960) identified coffinite, uranophane and soddyite in altered material from the Nordic Mine. Yellow secondary minerals may develop along water-bearing fractures after exposure in mine workings.

Pyrite is the commonest sulphide mineral; it usually constitutes some 6 to 10% of the ore and is concentrated in the matrix; only rarely is there indication of replacement or fracture filling in the quartz pebbles. Individual pyrite grains may be rounded ("buckshot" pyrite) or subhedral; alternatively, pyrite may occur in a massive habit, although spongy varieties are also present. Arnold (1954) suggested that the pyrite formed by sulphidization of detrital magnetite, and he described grains showing cores rich in leucoxene that he considered to have developed from ilmenite exsolved from the original magnetite. Bottrill (1971) and Robertson (1971) have suggested that the necessary sulphur may have been derived from adjacent Huronian volcanic rocks.

The postdepositional replacement of titaniferous magnetite has been confirmed by the studies of Theis (1979) and Robinson and Spooner (1982, 1984a, 1984b). This confirmation effectively counters the arguments of Davidson (1957), Simpson and Bowles (1981), and Clemmy (1981) which advocated that the lack of magnetite in the ore was a serious difficulty for the placer theory. Theis (1979) considered that the greater part of the pyrite was detrital and noted shapes and textures similar to those illustrated from the Witwatersrand Basin by authors such as Hallbauer (1981), Saager (1981), and Minter (1976). Pienaar (1963), in a trace-element study, could not distinguish between the pyrite of the ore beds and the pyrite found in other rocks of the Elliot Lake and Blind River districts. Davidson (1957) and Kimberley et al. (1980) have indicated that Co:Ni ratios of the pyrite are characteristic of igneous rocks. Ross (1981) suggested that the gold present in the ores was confined to the pyrite and was probably introduced in fluids derived from volcanism. Most authors have followed Roscoe (1969), who suggested that pyrite was polygenetic with detrital, authigenic, and introduced pyrite all being present. Robinson and Spooner (1982, 1984a, 1984b) considered the bulk of the pyrite to be authigenic with some detrital pyrite also being present. Overgrowths of authigenic pyrite on detrital pyrite are common.

As distinct from the Witwatersrand (Pretorius 1981) and Jacobina (Gross 1968; Villaça and Moura 1981) deposits, gold is a very minor component of the Blind River ores. Mining companies do not routinely assay for gold. However, Boyle (1979) indicated that, in the samples available to him, whole-rock gold content was typically 0.09 ppm Au and that this gold was largely contained within the pyrite where a focussed analysis on individual grains yielded 0.64 ppm Au. Ross (1981) obtained values in the range of from 5 to 987 ppb Au with individual "ore reefs" averaging from 73 to 117 ppb Au and one reworked "ore reef" at the base of the Ramsay Lake Formation conglomerate attaining 400 ppb Au. A wide variety of other sulphides and heavy minerals has been identified but these occur in very small quantities (Roscoe 1969). Mossman and Harron (1983) and Mossman (1987) have listed the known gold occurrences occurring within the Huronian Supergroup.

The predominant ore mineral at the Pronto Mine and in the Quirke Mine's A Reef (Theis 1979) is "brannerite", first described from the area by Nuffield (1954). "Brannerite" is typically found as heterogeneous assemblages of bladed rutile and quartz surrounded by a uranium oxide and rare earth oxides; the whole approximates UTi_2O_6 in composition and is regarded by Nuffield (1954) as a metamict product of crystalline brannerite broken down by radioactive bombardment. Pienaar (1963) described the aggregates as a mixture of quartz and a two-phase uranium-titanium compound. Ramdohr (1957) suggested that the "brannerite" was formed by the "Pronto-reaction":

$$UO_2 + (2\text{-}3)TiO_2 = UTi_{2\text{-}3}O_{6\text{-}8}$$

Ramdohr (1957) held that this reaction took place during metamorphism.

Ferris and Ruud (1971) and Theis (1979) (*see* below) have concluded that the Blind River area's "brannerite" and the brannerite, or uraniferous leucoxene, of the Witwatersrand Basin formed at low temperatures during diagenesis as a result of uranium migrating from uraninite to decomposing ilmenite.

Uraninite is the second most important mineral at the Pronto Mine and is apparently the most important in both the Nordic zone and in the C Reef at Quirke Mine (Theis 1979). Generally it occurs as black, rounded to subhedral grains approximately 1 mm across. Fresh uraninite typically has 6.5% ThO_2 and a U:Th ratio of about 10, which indicates derivation from a pegmatitic source (Roscoe 1969; Robertson 1968, 1976; Theis 1979). Clastic grains of uraninite show a crude hydraulic equivalence to surrounding rock fragments. In many localities, graded foreset laminations are found, clearly indicating water transport and deposition. Ferris and Ruud (1971) and Theis (1979), from microscope and/or microprobe studies, have demonstrated that the margins of uraninite grains have been depleted in U. They concluded that the "brannerite" formed by dissolution of U in uraninite followed by redeposition on nearby Ti minerals during diagenesis.

Roscoe (1969) compiled the available U and Pb isotope data on the Blind River area ores and concluded that the data are compatible with a primary-age determination of about 2500 Ma for the ore-forming minerals. Episodic introduction of uranium mineralization was not indicated. Subsequent detailed studies have confirmed a Neoarchean age (2500±50 Ma) for the uraninite grains (Meddaugh and Holland 1981; Robinson and Spooner 1984a, 1984b).

Early descriptions of thucholite from Blind River area ore were given by Robertson (1968, 1970). Two types were recognized: disseminations and thin layers within ore beds; and globules in postore fractures along with postore sulphides and, rarely, palygorskite (described as pilolite by Kaiman and Horwood (1976)). Ruzicka and Steacy (1976) and Ruzicka (1981) have collected and documented banded thucholite from the Elliot Lake area and have endorsed a biogenic origin for it; banded thucholite probably represents former algal mats which also trapped fragmental uraninite. Willingham et al. (1985) have identified kerogens in material collected by Ruzicka, confirming the biogenic origin.

Mossman and co-workers have documented the presence of gold with kerogen in the uraniferous rocks of the Huronian Supergroup (Mossman and Dyer 1985; Dyer et al. 1988; Willingham et al. 1985). Kaiman and Horwood (1976) concentrated on the globular variety of thucholite occurring in fractures and suggested that such material formed by polymerization of diesel fumes from mining operations; however, most mine geologists now discount a postmining formation of such thucholite.

Roscoe (1959a, 1959b) described the monazites from the Blind River area ores and has pointed out that monazite can contain considerable uranium and is, therefore, one of the ore minerals. Grains are normally rounded to subangular and less than 0.3 mm in diameter. Theis (1979) provided detailed accounts of several "ore reefs" of the Quirke and Denison mines.

PRODUCTION

By December 1989, the Blind River–Elliot Lake mining camp had produced some 135 450 t uranium for an average recovered grade of 0.09% U (2.1 pounds U_3O_8 per ton). A part of this production includes minor amounts recovered from mine waters, and more recently from underground-leaching operations and recycled raffinate from the uranium refinery in Blind River. An additional 752 t were recovered by leaching of broken ore, both underground and on surface, at Agnew Lake. During 1990 recovered grade was approximately 1.75 pounds U_3O_8 per ton reflecting the mining of lower-grade ore as depletion of the ore bodies continued.

Minor amounts of thorium and yttrium—the details of which are not available to the public realm—have also been produced. In mid-1968, a consortium started producing yttrium at a plant adjacent to the Denison Mill.

Maximum mineral production from the mining camp was obtained in 1959 when some 9345 t uranium were produced.

RESERVES AND RESOURCES

Given the bedded nature of the deposits and the extent of diamond drilling, the resources of the mining camp have been known since the late 1950s and were generally quoted as approximately 400 000 to 500 000 t uranium (Griffith and Roscoe 1964). Changes have been made to this resource estimate, reflecting exploration, production, mine development and economic factors.

The most detailed published breakdown on Ontario's uranium resources (predominantly the Elliot Lake mining camp) is that given by the Porter Commission (1978): Ontario's total inventory in detail includes a measured 49 000 t uranium, an indicated 75 000 t uranium, an inferred 246 000 t uranium, and a prognosticated 182 000 t uranium at prices below $156/kg U (1978). These 1978 resource estimates also contained 100 000 t of recoverable ThO_2.

Origin of the Uranium Deposits of Blind River Type

Uranium and thorium mineralization occurs at a number of localities throughout the world in quartz-pebble conglomerates bearing appreciable pyrite and—especially at the Witwatersrand Basin in South Africa—gold in the matrix. The origin of these conglomerates has been much debated. The similarity of the deposits at Blind River to one or another of several of the well-known deposits—at the Witwatersrand Basin, South Africa; at Jacobina and the Quadrilátero Ferrífero, Brazil; minor occurrences in the Fortescue Group in the Pilbara area of Western Australia; and at unspecified localities in Russia and China—have been pointed out in Davidson (1957), Gross (1968), Robertson (1974), Armstrong (1981), Button and Adams (1981), Villaça and Moura (1981), and Derry (1986).

Earlier workers, such as Davidson (1957), Heinrich (1958) and Joubin (1960), proposed a hydrothermal origin for the ores, emphasizing the lack of uranium in Phanerozoic and modern placers. They cited the supposedly high U:Th ratios, the high Ti:Fe ratios, and the association of Ti, Co, Ni, Th, and U in a deposit carrying brannerite, uraninite, and pyrite (±gold) as characteristic evidence of a hydrothermal origin. Holmes (1957), D.S. Robertson (1962, 1974), Pienaar (1963), Robertson (1968, 1986), Roscoe (1969, 1981), Theis (1979) and Ruzicka (1981, 1988, 1989) have strongly supported a placer origin for the uranium ores with modification by later events based on the following clues: close spatial relationships of ore minerals to sedimentary structures; local reworking at younger disconformities; textural and mineralogic features which can be matched with those of an Archean provenance; the evidence provided by observed isotopic ratios; and no evidence of there having been any hydrothermal introduction from later diabase and/or granite. In the mines, diabase clearly cuts the ore beds, and locally there is albitic, chloritic or carbonate alteration of the ores (Robertson 1968, 1970).

Many authors—for example, Pienaar (1963), Roscoe (1969, 1973, 1981), and Robertson (1974)—have suggested that the regolith and the presence of uranium and pyrite in drab-coloured rocks implies a reducing atmosphere. Robertson (1961, 1971, 1976, 1977, 1981, 1986, 1989) has pointed out that even if limited oxygen was available, the rapid transportation and deposition of the uranium-bearing detritus under conditions in which glaciation was common would severely limit oxidation and solution of uranium.

The overall distribution of beds and the relationship of thickness and grades are more consistently explained by a modified placer hypothesis, first proposed by Holmes (1957), which Robertson (1961, 1971, 1976, 1977, 1981, 1986, 1989) has long supported. Robertson (1989) has published a genetic model for the Blind River ore deposits.

This modified placer hypothesis proposes that the Blind River ore deposits were derived from a cratonized Archean terrane positioned to the north and northwest—a terrane which would have included uraniferous granitic rocks with or without uraninite-bearing pegmatites. This hypothesis further proposes that the ore's mineral grains were transported by rapidly moving water and deposited in topographic depressions controlled by pre–Huronian Supergroup geology and erosion. Transport and deposition would have occurred under (cold?) oxygen-deficient conditions—but not necessarily in an atmosphere devoid of oxygen—approximately 2500 million years ago. Mafic volcanic rocks were simultaneously accumulating in the basin and may have provided a sulphur source to convert magnetite to pyrite, although much of the pyrite was probably derived from the cratonized Archean terrane. The uranium deposits were later subjected to diagenesis and to minor alteration during subsequent intrusive, metamorphic, and tectonic events, but they were not subjected to either extensive erosion or leaching. The modified placer hypothesis has been recently endorsed by Robinson and Spooner (1982, 1984a, 1984b) but with doubts expressed on the evidence for a reducing atmosphere.

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Chapter 15

The Sudbury Structure

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Abstract

The 1850 million-year-old Sudbury Structure, which is host to significant nickel-copper deposits, is located northwest of the Grenville Province, at the present boundary of the Superior Province and the Southern Province. It consists of the following:

1. **the Sudbury Igneous Complex, which comprises norite, quartz gabbro, granophyre and an inclusion-rich quartz dioritic unit called the Sublayer**
2. **the Sudbury Basin, which is occupied by the heterolithic breccias, mudstones and wackes of the Whitewater Group**
3. **brecciated footwall rocks, which occur around the Sudbury Igneous Complex**

The origin of the Sudbury Structure is controversial. The presence of heterolithic breccias in the Sudbury Basin and around the Sudbury Igneous Complex, and of shock metamorphic features in these breccias and the footwall rocks, is indicative of a catastrophic event, either the impact of a meteorite or an endogenic explosion.

The Sudbury Structure has been deformed by northwesterly directed thrusting and subjected to regional metamorphism during the Penokean Orogeny.

Data from gravity and magnetic surveys suggest that the Sudbury Structure is underlain by a large, dense mass, probably of mafic intrusive rocks, which extends beyond the present limits of the Sudbury Igneous Complex. However, preliminary results from deep seismic surveys appear not to confirm the presence of mafic intrusive rocks at depth.

Nickel and copper, along with by-products cobalt, platinum group metals, gold, silver, selenium, tellurium, sulphuric acid, liquid sulphur dioxide and slag are produced from Sudbury ores. The mineralization occurs in the Sublayer, which is at the lower contact of the Sudbury Igneous Complex and within the offset dikes. It also occurs in the Footwall Breccia and in the footwall rocks themselves. The ores are attributed to separation of sulphides from a sulphur-rich mafic magma.

INTRODUCTION

The Paleoproterozoic Sudbury Structure is an unique geological feature that is known worldwide for its significant nickel-copper ore deposits and its contentious origin. The Sudbury Structure is located in south-central Ontario, northwest of the Grenville Province, at the present boundary of the Archean Superior Province with the Proterozoic Southern Province[1].

The Sudbury Structure is defined by the norites, quartz gabbro, and granophyres of the Sudbury Igneous Complex; the heterolithic breccias, mudstones, siltstones and wackes of the Whitewater Group which occupy the Sudbury Basin; and the brecciated footwall rocks around the Sudbury Igneous Complex. In plan view, the Sudbury Igneous Complex is elliptical in shape and approximately 27 by 60 km in size (Figure 15.1). Two types of breccia, which are related to the genesis of the Sudbury Structure, occur in the footwall of the structure; the heterolithic Footwall Breccia, which lies at the lower contact of the Sudbury Igneous Complex (*see* Figure 15.1), and the Sudbury Breccia (pseudotachylytes), which is very widespread and occurs up to 80 km north of the Complex. The footwall rocks include Archean gneisses and granitic and mafic igneous rocks to the north, west and east, and metavolcanic and metasedimentary rocks of the Paleoproterozoic Huronian Supergroup to the south of the Complex. The Paleoproterozoic Creighton and Murray granitic plutons intrude the Huronian sequence along the south-central margin of the Sudbury Igneous Complex and locally form the footwall rocks of the Complex in this area.

The Whitewater Group consists of impact-generated and/or volcanogenic breccias of the Onaping Formation, the pelagic sedimentary rocks of the Onwatin Formation and turbiditic wackes of the Chelmsford Formation. The group is contained in the central depression of the Sudbury Structure known as the Sudbury Basin.

The Sudbury Igneous Complex lies structurally below the Whitewater Group. The lowermost unit of the Complex is the Sublayer. It is known as the Contact Sublayer where it intrudes along the contact between the Main Mass of the Complex and the footwall rocks, and it is known as the Offset Sublayer where it forms the offset dikes in the surrounding rocks. The Sublayer contains many of the nickel-copper ore bodies in the Sudbury area; but ore bodies also occur in the footwall rocks and the Footwall Breccia.

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[1] *The Southern Province is here defined, as is customary in the geological literature, as that region of the Canadian Shield underlain by rocks of the Huronian Supergroup and other Proterozoic rocks.*

The offset dikes and the Contact Sublayer range from gabbro to quartz diorite and feature a multitude of inclusions derived from the footwall. Mafic to ultramafic inclusions also occur and are commonly associated with Sublayer orebodies. They are possibly derived from the lowermost, unexplored parts of the Sudbury Igneous Complex (Naldrett 1984c).

The origin of the Sudbury Structure, the Sudbury Igneous Complex and associated ores has long been a subject of controversy. The heterolithic Footwall Breccia, the pseudotachylytes (Sudbury Breccia) around the Complex, the heterolithic breccias of the Onaping Formation, and shock metamorphic features in the various breccias and the footwall rocks are, collectively, indications of a catastrophic event. For one school of thought, this catastrophic event was an endogenic volcanic explosion. Another school explains these features as the result of the impact of a large meteorite.

The rocks of the Sudbury Igneous Complex were commonly considered as magmatic units (Barlow 1904, 1906; Coleman et al. 1929; Walker 1935; and many authors since) whose intrusion possibly was triggered by the catastrophic "Sudbury Event" (Dietz 1964; Peredery and Morrison 1984; Dressler et al. 1987). Dence (1972) considered portions of the micropegmatite (granophyre) of the Sudbury Igneous Complex to represent an impact melt. More recently, Faggart et al. (1985) and Stöffler et al. (1989), interpreted all the units of the Complex to be impact melts, based on geochemical and planetological considerations, including impact models.

Likewise, diametrically opposed views have been expressed on the origin of the Sudbury ores. Conventionally, the sulphide ores were thought to be of a magmatic (Bell 1891; Naldrett 1984a) or hydrothermal (Dickson 1903; Wandke and Hoffman 1924) origin, whereas Dietz (1972)

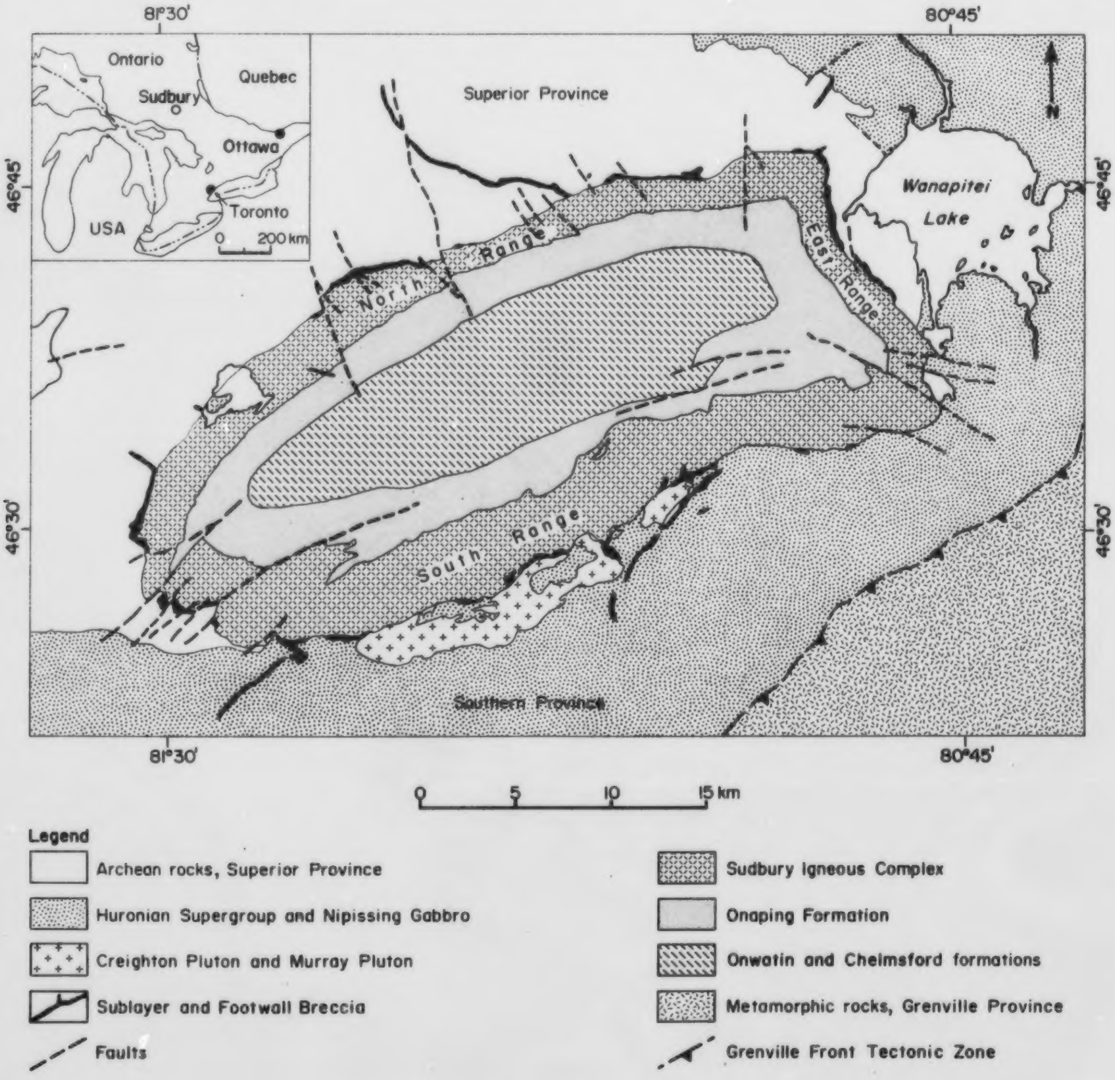
81°30'
80°45'
46°45'
46°30'
Ontario
Quebec
Sudbury
Ottawa
Toronto
USA
0 200 km
Superior Province
N
Wanapitei Lake
North Range
East Range
South Range
Southern Province
0 5 10 15 km
Legend
Archean rocks, Superior Province
Huronian Supergroup and Nipissing Gabbro
Creighton Pluton and Murray Pluton
Sublayer and Footwall Breccia
Faults
Sudbury Igneous Complex
Onaping Formation
Onwatin and Chelmsford formations
Metamorphic rocks, Grenville Province
Grenville Front Tectonic Zone

Figure 15.1. General geology of the Sudbury Structure.

proposed a derivation from an impacting bolide. Naldrett (1984a) advocated a magmatic origin for the mineralization in which sulphide precipitation was triggered by the contamination of a mafic magma through the assimilation of silica-rich country rock.

In the following sections, the rock units and geological events are described in a chronological sequence. A short historical overview on the development of geological concepts and interpretations precedes most sections of this review. The chapter concludes with a summary of current thinking on the origin of the Sudbury Structure.

Much of this overview is based on a detailed volume entitled *The Geology and Ore Deposits of the Sudbury Structure* (Pye et al. 1984). Many figures in this chapter are taken from that volume. Review papers on the Sudbury Structure and the regional geology of the Sudbury area are by Card et al. (1984) and Dressler (1984a). Gupta et al. (1984) dealt with the geophysical characteristics of the Sudbury Structure, and Giblin (1984a) presented a glossary of Sudbury geology terms.

DISCOVERIES AND MINING

In 1856, a surveyor by the name of Salter discovered sulphide minerals at a location near that which later became the Creighton Mine and reported his findings to Alexander Murray of the Geological Survey of Canada (Giblin 1984b). Murray (1857) described the occurrence in a government report. In 1883, commonly cited as the year of the first discovery of the nickel-copper orebodies in the Sudbury area, blacksmith T. Flanaghan noticed a gossan on the right-of-way for the Canadian Pacific Railway line which was being constructed at this time. T. Murray developed this occurrence which became the Murray Mine. In 1886, the first ore in the Sudbury Mining Camp was produced by the Canadian Copper Company at the Copper Cliff Mine. A smelter began operation in the same year.

The Sudbury ores were mined solely for their copper content until it was recognized that the ores contained appreciable nickel. However, only after the creation of a market, for example in naval armour plates for the US Navy, did nickel become a marketable commodity early in this century.

The major companies involved in the development of the Sudbury mines early in this century were the Mond Nickel Company, formed in 1900; the International Nickel Company, incorporated in 1902; and the British American Nickel Corporation Limited. In 1929, the Mond Nickel Company was absorbed by the International Nickel Company. Falconbridge Nickel Mines Limited was founded in 1928 and commenced smelting in 1930.

Presently, Falconbridge Limited and Inco Limited are the sole producers in the Sudbury area. In 1974, the peak of nickel production was reached when 209 000 tonnes were produced. In 1988, the 2 companies produced 128 558 tonnes of nickel (21% of the nickel consumption of the western world) having a value of $1 953 566 137. By-products of nickel-copper production include cobalt, the platinum group metals, gold, silver, selenium, tellurium, sulphuric acid, liquid sulphur dioxide, and slag for road construction. In 1989, due to the high price of nickel, the mineral production from the Sudbury Structure had a value of about $2 639 000 000. This constituted approximately 37% of the total value of Ontario's mineral production (including non-metal minerals, aggregates, and fuels) and about 13% of the total value of Canada's production. For comparison, the total value of the 1988 production of gold in Ontario was about $1 006 000 000 or about 5% of the total value of mineral production in Canada.

GENERAL GEOLOGY

Introduction

The Sudbury Structure is perhaps the most intensively studied geological feature in Ontario and a review of the literature describing it has to be selective. Nevertheless, this chapter gives an overview of the present knowledge of the structure in an order reflecting the sequence of geological events as it is presently understood.

The catastrophic Sudbury Event, an endogenic explosion (*see* Muir 1984 for the most recent review and references) or meteorite impact (*see* Dressler et al. 1987 for the most recent review and references), is believed to have caused extensive brecciation and deformation and possibly, as advocated by proponents of the impact hypothesis, large-scale tectonism of the country rocks of the Sudbury Structure. Rock units formed by this event include pseudotachylytes (Sudbury Breccia), Footwall Breccia and heterolithic breccias of the Onaping Formation. Shock metamorphic features occur in these breccias and the footwall rocks of the Sudbury Igneous Complex, but not in the units of the Complex and the rocks of the upper 2 formations of the Whitewater Group (*see* below). An impact scenario requires that tectonism, brecciation and shock metamorphism occurred almost simultaneously, whereas in an endogenic event simultaneity is not required.

The Whitewater Group consists of the heterolithic breccias of the Onaping Formation at its base, overlain by the mudstones of the Onwatin Formation and the wackes of the Chelmsford Formation. The 2 upper formations are not directly related to the Sudbury Event.

The Sudbury Igneous Complex is overlain by, and has intrusive contacts with, breccias of the Onaping Formation. The Complex is, therefore, commonly believed to have been emplaced after the Sudbury Event. This belief, however, has recently been contested by some proponents of the impact hypothesis (Faggart et al. 1985; Stöffler et al. 1989). The age relationship of the Onwatin and Chelmsford formations with respect to the Sudbury Igneous Complex is not known (*see* below).

The Sudbury Igneous Complex consists of several units. Norites, quartz gabbro and granophyres make up its "Main Mass" and the Sublayer forms its base. Several gabbroic and inclusion-bearing dikes intrude rocks of the Whitewater Group. If these are related to, and contemporaneous with, the Sudbury Igneous Complex, they would indicate an age for all of the Whitewater Group that predates the emplacement of the Complex.

The Sudbury orebodies are described in the last part of this section on the general geology of the Sudbury Structure. Table 15.1 summarizes the lithostratigraphy and geochronology of the rock units of the Sudbury area. The Sudbury Event itself cannot be dated directly, but its age is probably close to 1850 Ma (Krogh et al. 1984), the age of the Sudbury Igneous Complex.

Tectonism, Brecciation and Deformation of the Footwall Rocks of the Sudbury Structure

TECTONISM—OVERTURNED COUNTRY ROCKS NEAR THE SOUTH RANGE STRUCTURE

In several places (Drury, Denison and Falconbridge townships), the South Range (*see* Figure 15.1) supracrustal footwall rocks of the Huronian Supergroup (Bennett et al., this volume) dip vertically or are overturned and north dipping where they are within 100 to 750 m of the Sudbury Igneous Complex (Card 1968; Dressler 1984c, 1987). Farther away from the Sudbury Igneous Complex, dips are to the south and, in general, decrease in magnitude from 90° to about 35° (Dressler 1987). Along much of the contact with the Complex, the South Range footwall rocks between Denison and Falconbridge townships consist mainly of metavolcanic rocks with equivocal top indicators, and granites.

The overturned rim of the structure is possibly related to the Sudbury Event; however, several tectonic episodes appear to have affected the rocks south of the Sudbury Igneous Complex. Therefore, at present, the timing of the overturning of the country rocks in the chronology of the deformation is not clearly understood. In the North and East ranges (*see* Figure 15.1), the country rocks near the Sudbury Igneous Complex consist mainly of granitic rocks and gneisses and, for these reasons, no evidence exists for or against overturning. An overturned rim of the Sudbury Structure that was formed before the deposition of the rocks of the Whitewater Group, and before the emplacement of the Sudbury Igneous Complex, would support an impact origin for the Structure, as overturned collar rocks have been observed in other impact structures (Shoemaker 1963; Offield and Pohn 1977; Daly 1947).

BRECCIAS IN THE FOOTWALL ROCKS

Sudbury Breccia and Footwall Breccia are some of the most intriguing rock types related to the Sudbury Structure and its origin. Their origin remains a subject of considerable debate.

Sudbury Breccia

Sudbury Breccias are pseudotachylytes, very similar to those described by Shand (1916) from the type location of the Vredefort Structure in South Africa. They occur peripheral to the Sudbury Igneous Complex in rocks of the Southern and Superior provinces, up to a distance of about 80 km from the Sudbury Igneous Complex. Previously known as Sudbury-type breccia, Levack Breccia, or Common Sudbury Breccia, the rocks have been studied extensively (Fairbairn and Robson 1942, 1943; Speers 1957; Card 1978a; Dupuis et al. 1982; Dressler 1984c; Brockmeyer 1986; and Müller–Mohr 1988).

Sudbury Breccia bodies (Figures 15.2a and 15.2b) occur within all footwall rocks predating the Sudbury Event, and are most abundant in a zone 5 to 10 km wide at the contact with the Sudbury Igneous Complex. Field mapping suggests the presence of 2 or more discontinuous zones of increased brecciation about 20 to 25 km, and 80 km, from the Sudbury Igneous Complex (Dressler 1984c; Peredery and Morrison 1984). The breccias form irregular and tabular bodies ranging from a few millimetres thick to breccia zones 0.5 by 11 km in size in the well-exposed rocks of the Huronian Supergroup south of the Complex. North of the Complex, where outcrops are less plentiful, the largest known breccia body is 100 to 200 m by 30 to 50 m in areal extent. Most breccia dikes dip vertically or steeply and have no obvious preferred orientation with respect to the present shape of the Sudbury Structure (Dressler 1984c). Sudbury Breccia dikes cutting across other Sudbury Breccia dikes, and fragments of Sudbury Breccia within Sudbury Breccia have been observed.

The contacts between the breccia bodies and their host rocks are commonly sharp. Transitional contacts are scarce (Figure 15.3). Large fragments within the breccia are commonly subrounded to rounded, the smaller ones are commonly angular. In general, fragments are of the same rock type as their immediate host rocks. "Exotic" fragments (i.e., fragments from sources not exposed nearby) may occur in subordinate amounts, more so in larger bodies than in smaller ones.

The matrix (Figure 15.4) of the Sudbury Breccia is grey or black and commonly consists of a microscopic to submicroscopic, weakly recrystallized, massive rock (Dressler 1984c). Amygdules, microliths and igneous textures have been observed in a few places. Cataclastic flow banding also occurs but is not common. Regional metamorphism, particularly south of the Sudbury Igneous Complex, and a 1.5 km wide contact metamorphic aureole related to the Sudbury Igneous Complex have obliterated many microscopic features.

The Sudbury Breccia is considered to have formed by sudden, explosive brittle failure leading to a violent milling and crushing process that involved little or no melting. Fusion, however, has occurred locally and is indicated in places by amygdaloidal or non-amygdaloidal melt matrices, microliths and igneous textures (Dressler 1984c).

Several geochemical investigations (Speers 1957; Card 1978a; Dressler 1984c) suggested that the breccias were formed by an *in situ* process involving host rocks of the breccias and fragments within the breccias. Dupuis et al. (1982) published geochemical results which they interpreted as evidence for an involvement of alkali-rich fluids and alkalic

Table 15.1. Lithostratigraphy and Geochronology, Sudbury Area[1].

SOUTHERN PROVINCE		SUPERIOR PROVINCE
Footwall Rocks	**Sudbury Structure**	
PROTEROZOIC		
Olivine Diabase (1250 Ma)	Olivine Diabase (1250 Ma)	Olivine Diabase (1250 Ma)
Intrusive Contact	Intrusive Contact	Intrusive Contact
	Trap Dikes[2]	
	Intrusive Contact	
	Sudbury Igneous Complex Sublayer[2] Intrusive Contact Main Mass (1849 Ma) Gabbro[2,3]	
	Whitewater Group Chelmsford Formation[2] Onwatin Formation[2]	
	SUDBURY EVENT Onaping Formation[4] Sudbury Breccia Footwall Breccia shock metamorphism	
Nipissing Intrusive Rocks (2220 Ma)		
Intrusive Contact		
Creighton (2333^{+32}_{-22} Ma)[6] and Murray (2388^{+20}_{-13} Ma)[1] Granites		
Intrusive Contact		
Huronian Supergroup Copper Cliff Formation: rhyolite (2450^{+25}_{-10} Ma)		
ARCHEAN		
Archean Basement		Pegmatite (2647±2 Ma) Levack Gneiss (2711±7 Ma) U-Pb; 1950 Rb-Sr[5]

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[1] *U-Pb, for details see Krogh et al. 1984.*

[2] *Absolute age unknown.*

[3] *Gabbro in interior of Sudbury Basin, possibly related to Main Mass or Sublayer of the Sudbury Igneous Complex; intrusive contacts with all formations of the Whitewater Group.*

[4] *Lower intercept age of inherited Archean (2711 Ma), shock metamorphic zircons in clasts of the Onaping Formation are 1836±14 Ma.*

[5] *U-Pb, Andrews et al. 1986.*

[6] *U-Pb, Frarey et al. 1982.*

magmatism in the formation of the Sudbury Breccias. This interpretation has been disputed by Dressler (1984c).

Footwall Breccia

The Footwall Breccia has been described under various names such as "late granite breccia" and "leucocratic breccia" (Langford 1960; Souch et al. 1969; Greenman 1970; Pattison 1979; Muir 1981, 1983; Dressler 1984c, Lakomy 1986, 1989). It contains much of the nickel-copper mineralization of the Sudbury Structure, particularly in the North Range. As "leucocratic breccia" it had been included as part of the Sublayer by Pattison (1979).

The Footwall Breccia is most common in the North and East ranges where it occurs as sheets and discontinuous bodies up to 150 m thick, parallel to the lower contact of the Sudbury Igneous Complex. Also observed are small Footwall Breccia dikes in the Archean gneisses north of the complex. They are up to 1 m wide and occur up to 250 m from the

Figure 15.2. Sudbury Breccia: a) breccia body in Creighton Pluton, b) steeply dipping breccia dike in Creighton Pluton.

main breccia bodies. Footwall Breccia inclusions in both the Contact Sublayer and the Offset Sublayer, and the in the norite of the Complex, have also been observed. Contacts between the breccia and the overlying Sudbury Igneous Complex are commonly sharp; and those with the underlying footwall rocks are gradational across a "megabreccia" zone (Pattison 1979) up to 150 m wide.

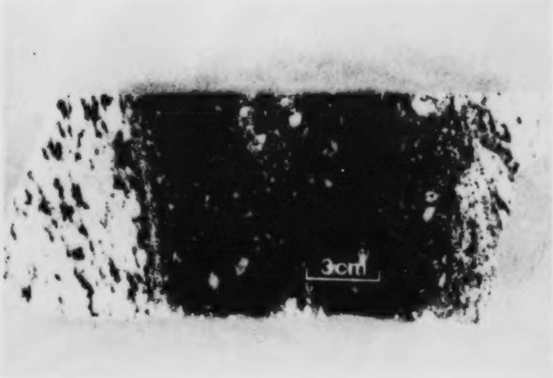


Figure 15.3. Sudbury Breccia with transitional contacts. The host rock is granite. Matrix is fine grained and black in central part (75% of width) of breccia dike, and grey and somewhat coarser grained at both borders (25% of width). All clasts are granite. Creighton Pluton.

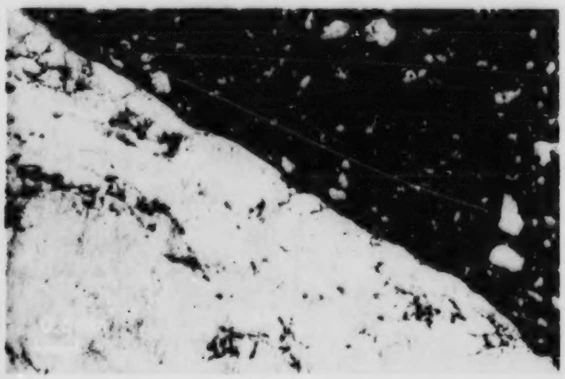


Figure 15.4. Contact of Sudbury Breccia matrix with granitic host rock. North of Sudbury Igneous Complex. Plane polarized light.

The Footwall Breccia is heterolithic and consists of angular to subrounded fragments of a wide range of sizes (Figure 15.5). Most rock fragments are derived locally, but exotic fragments also occur (Dressler 1984c). The matrix is commonly light coloured and has been characterized by Pattison (1979) as "mosaic granoblastic metamorphic" in texture. This description, however, does not encompass all microscopic features. Dressler (1984c) recognized features indicating shock metamorphic deformation of rock-forming minerals, and distinct granoblastic and granophyric textures in the matrix. These granoblastic and granophyric features are considered to be contact metamorphic in origin (Dressler 1984c). Some textures suggest that incipient melting has occurred close to the contact between the Footwall Breccia and the Sudbury Igneous Complex (Dressler 1984c; Lakomy 1989). Migration of matter within the breccia due to the contact metamorphic overprint was minor. However, eutectic, commonly granophyric, melt fractions in the matrix occur in many places, and have rarely coalesced into granitoid veins and dikes (Dressler 1984c).

It is the view of the authors that the Footwall Breccia probably consisted of a parautochthonous mass of crushed and shock-metamorphosed rocks and hence originally was similar to Sudbury Breccia. The Footwall Breccia formed discontinuous masses along the walls of the Sudbury Basin before the deposition of the heterolithic breccias of the Onaping Formation and the intrusion of the Sudbury Igneous Complex. These walls are believed to have been the original endogenic-volcanic or meteorite impact crater walls. Before the intrusion of the Sudbury Igneous Complex, the Footwall Breccia was in contact with the heterolithic breccias of the lower Onaping Formation. In the South Range, because of uplift and subsequent erosion, a lower section of the Sudbury Igneous Complex and its footwall rocks is exposed (Naldrett et al. 1970). Coupled with the present abundance of Footwall Breccia in the North Range relative to the South Range, this suggests that Footwall Breccia bodies were more common along the upper than the lower crater walls.

SHOCK METAMORPHISM

Shock metamorphism refers to changes in rocks and minerals resulting from the passage of high-pressure shock waves. Shock metamorphic features have been observed in the footwall of the Sudbury Igneous Complex (Dence 1972; Guy-Bray et al. 1966; Dressler 1984c) and in the breccias of the Onaping Formation (French 1968, 1970, 1972; Peredery 1972a, 1972b; Muir and Peredery 1984). Their presence has been acknowledged by all recent researchers investigating the Sudbury Structure, and is not contested by proponents of an endogenic origin for the Sudbury Structure.

Shatter cones occur in all rocks around the Sudbury Igneous Complex that predate 1850 Ma. They have been observed as far as 17 km from the Complex (Guy-Bray et al. 1966; Dressler 1984c). Microscopic shock metamorphic features in the footwall consist of planar elements in quartz and feldspar (Figure 15.6), kink bands in mica and extreme fracturing in garnet. Dressler (1984c) investigated the zoning of shock metamorphic features in the North Range footwall (Figure 15.7). Planar elements in quartz have been recognized up to 5.5 km north of the Sudbury Igneous Complex, but are present as far away as 8 km, as shown by Dence (1972). South of the Complex, microscopic shock metamorphic features are not observed due to regional metamorphism associated with Penokean orogenic events.

Shock metamorphic, planar elements are also present in the heterolithic breccias of the Onaping Formation. These breccias also feature a variety of inhomogeneous and homogeneous recrystallized glasses (Muir and Peredery 1984) that are variously interpreted as diaplectic glasses (Peredery 1972a, 1972b; Peredery and Morrison 1984), which formed when the rocks were subjected to shock pressures greater than approximately 30 GPa, or as recrystallized volcanic glasses (Muir 1984).

Figure 15.5. Footwall Breccia. All fragments are locally derived tonalitic and mafic gneisses set in a strongly recrystallized matrix of mineral and rock fragments. Fluidal textures are indicative of incipient melting. Near Coleman Mine, Levack Township.

Figure 15.6. Planar features in potassic feldspar in granite; North Range. Plane polarized light.

The Whitewater Group

ONAPING FORMATION

The Onaping Formation structurally overlies the Sudbury Igneous Complex and consists of enigmatic breccias and igneous-textured rocks of contentious origin, which extend the full length and width of the Sudbury Basin, about 53 by 17 km (*see* Figure 15.1).

The Onaping Formation is the lowermost and most voluminous formation of the Whitewater Group. Volume estimates range from 530 km³ (Coleman 1905), to 1250 km³ (Williams 1957), and to 1670 km³ (Stevenson 1972). At its base, the formation is intruded by granophyres of the Sudbury Igneous Complex. The formation grades into the overlying fine siltstones and claystones of the Onwatin Formation.

The Onaping Formation is subdivided into 3 major units; from the base upward they are the Basal Member (Figure 15.8), the Gray Member (Figure 15.9), and the Black Member (Figure 15.10). The 2 upper members are

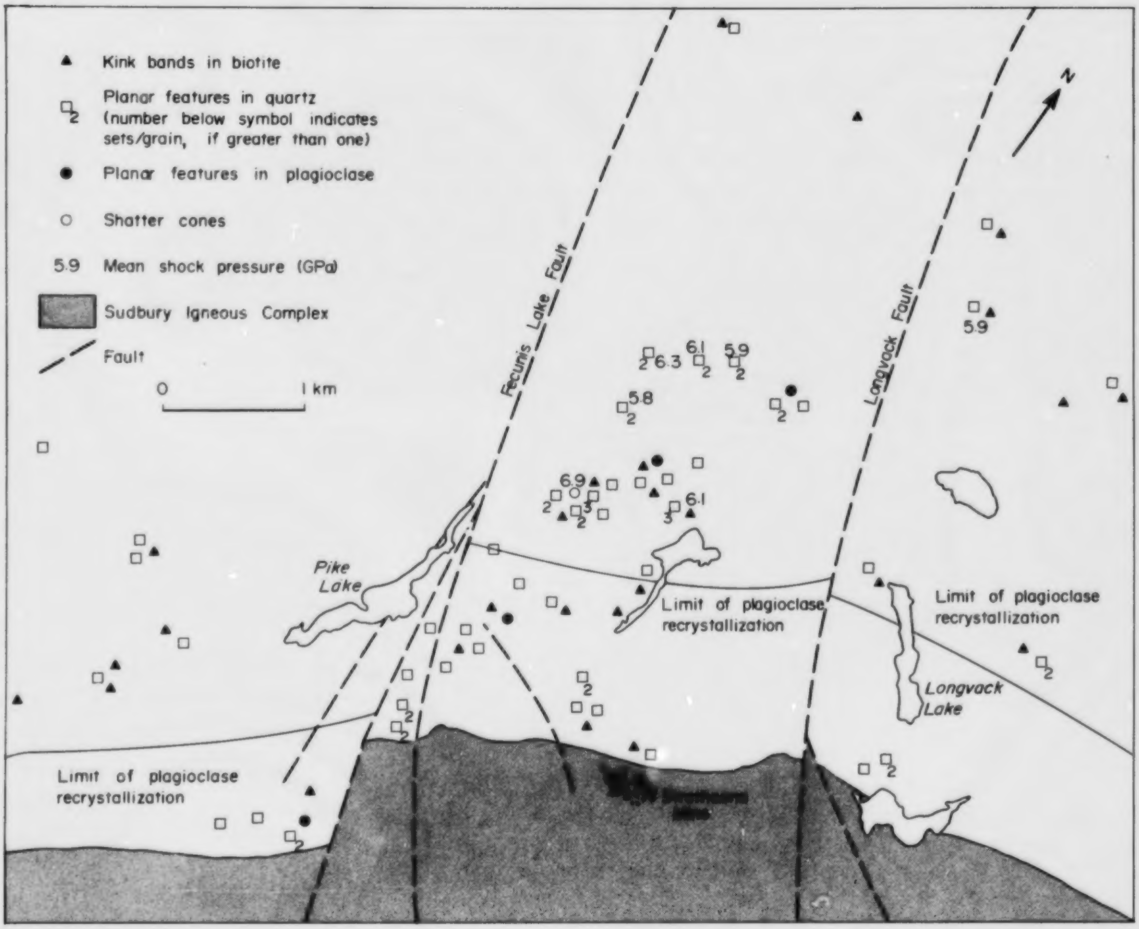

▲ Kink bands in biotite
□₂ Planar features in quartz (number below symbol indicates sets/grain, if greater than one)
● Planar features in plagioclase
○ Shatter cones
5.9 Mean shock pressure (GPa)
▨ Sudbury Igneous Complex
Fault
0
1 km
N
Fecunis Lake Fault
Longvack Fault
Pike Lake
Longvack Lake
Limit of plagioclase recrystallization
Limit of plagioclase recrystallization
Limit of plagioclase recrystallization
5.9
6.1
6.3
5.9
5.8
6.9
6.1
2
3

Figure 15.7. Zoning of shock metamorphic features in the footwall rocks of Levack Township. Shock pressures in GPa. (*from* Dressler 1984c, *calculation after* Robertson and Grieve 1977).

distinguished, as their names suggest, by colour. These members collectively comprise complexly layered, generally heterolithic breccias. Included in the formation are interlayered and discordant lenses of igneous- and fluidal-textured material known as Melt Bodies, which form a fourth unit.

The Onaping Formation has been described and interpreted to be the product of either: 1) volcanism (Bell 1893; Coleman 1905; Burrows and Rickaby 1930; Thomson 1957; Williams 1957; Speers 1957; Stevenson 1961a, 1961b, 1972; discussion by Muir *in* Muir and Peredery 1984; Muir 1984, 1986); 2) meteorite impact (French 1968; Dence 1972; Peredery 1972a, 1972b; discussion by Peredery *in* Muir and Peredery 1984; Peredery and Morrison 1984; Brockmeyer and Deutsch 1989; Stöffler et al. 1989); or 3) impact-induced volcanism (Dietz 1964; Thomson 1969; Muir 1982, 1983). The most recent and detailed description of the Onaping Formation was published by Muir and Peredery (1984), and this forms the basis for the following description.

Basal Member

The Basal Member (*see* Figure 15.8) is discontinuous. In the North and East ranges, the unit is generally less than 150 m thick, but locally up to 300 m thick. It comprises medium to coarse breccias consisting of fragments of Archean country rocks (particularly leucocratic granitoids), fragments of Huronian country rocks (particularly quartz arenite and arkose), and a fine-grained, commonly heterogeneous matrix composed of recrystallized country rock particles and/or igneous-textured minerals. Brockmeyer (1990) interprets the matrix of the Basal Member as igneous and the Basal Member as part of impact melt system also comprising the Sudbury Igneous Complex. Subunits of breccias, based on fragment size and/or fragment-to-matrix ratio, are locally distinguishable (Peredery 1972b; Muir 1983). These rock units have widely variable thicknesses, up to about 10 m, and are separated by gradational or sharp contacts. The contact between the Basal Member and the underlying Sudbury Igneous Complex is gradational in the case of "common granophyre", or sharp in the case of

Figure 15.8. Basal Member of the Onaping Formation (quartzite breccia, Stevenson 1961a). Angular to subrounded fragments of metamorphosed quartz arenite in a recrystallized quartzofeldspathic matrix. Hammer is 0.4 m long. Garson Township.

Figure 15.10. Black Member of the Onaping Formation. Small- to medium-sized fragments of recrystallized glass and country rocks in a matrix consisting of fragments of recrystallized glasses and country rocks. North Range.

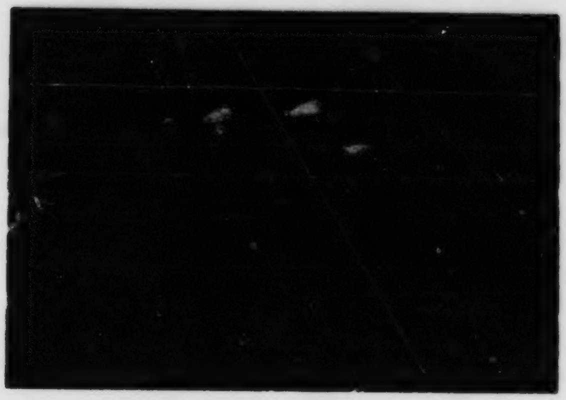


Figure 15.9. Gray Member of the Onaping Formation. Block-sized fragments of various country rocks, fluidal-textured material, and breccia within a matrix consisting of fragments of country rocks and recrystallized glasses. North Range.

"plagioclase-rich granophyre" (*see* section titled "Sudbury Igneous Complex"). The contact with the Gray Member is sharp.

In the South Range, the Basal Member has locally been tectonically transposed or faulted into place within the Gray Member, in which case it may appear much thicker. Here, fragments are mostly quartz arenite, with lesser arkose, conglomerate, feldspar-sericite schist and quartz (quartzite breccia of Stevenson 1961a; *see also* Figure 15.8). Fragments range from centimetre scale to blocks several tens of metres long. The matrix is generally recrystallized and moderately to strongly foliated.

At several localities in the North and South ranges, there is no distinct contact between the Basal Member and the adjacent Melt Bodies (Muir 1983; Brockmeyer and Deutsch 1989), which suggests that the 2 units may be genetically related. Locally, the Basal Member and Footwall Breccia are similar, and only the presence of the Sudbury Igneous Complex between the 2 breccia units permits a distinction to be made. It is not clear whether or not the 2 units were gradational into each other before the intrusion of the Sudbury Igneous Complex.

Gray Member

The Gray Member (*see* Figure 15.9) is a major, laterally continuous unit of breccia which ranges from 200 to 700 m thick. Its lower contact with the Basal Member, and locally with the Sudbury Igneous Complex, is sharp. Its upper contact with the Black Member is generally gradational, but locally sharp. The Gray Member locally comprises a number of distinct subunits, with contacts that are sharp or gradational at centimetre to metre scale.

Individual subunits, where well defined, range from one to several metres thick and display a lack of lateral continuity beyond several tens of metres, although some distinctive units can be traced for up to several kilometres (Stevenson 1961b; Muir 1982). Contacts range from straight to sinuous. Most units appear to dip towards the centre of the Sudbury Basin, but some dip away from it at about 40° to 80°.

The breccias of the Gray member consist of angular to rounded fragments of country rock, crystal fragments, recrystallized glass, fluidal-textured material and sulphides, which collectively range in size over at least 6 orders of magnitude, from less than 0.1 mm to perhaps a hundred metres (Figure 15.11). The country rock fragments are of various types derived from the Archean and Proterozoic country rock units, such as granitic rocks, gneisses, metavolcanic and metasedimentary rocks. Most individual crystals appear to be derived from comminuted country rocks, but some are euhedral to subhedral and appear similar to phenocrysts. Some lithic and crystal fragments display shock metamorphic features in quartz and plagioclase, as well as cryptocrystalline and devitrification textures in feldspar. The latter textures may represent altered, shock metamorphic (i.e., diaplectic) glasses (French 1968, 1972).

The recrystallized glass fragments range from heterogeneous to homogeneous. Massive, aphanitic, spherulitic, amygdaloidal and crystallite-bearing textures have been observed. Compositions range from predominantly felsic to predominantly mafic. Shapes range from well rounded to angular and shard-like. Some recrystallized glass fragments, generally within particular subunits of the Gray Member, are rimmed by axiolitic or segmented albite and/or potassic feldspar crystals.

The fragments of fluidal-textured material are aphanitic, and may display weakly to well-developed, undeformed to highly contorted flow banding, as well as amygdules, vesicles and spherules. Hence, it can be concluded that the material was likely originally in a glassy state when it cooled. The flow banding in some fragments is concordant with the fragment margins, and in others is truncated by fragment boundaries. Based on the presence of discontinuous, compositionally different layers with flame-like terminations, some flow-banded, fluidal-textured material resembles densely welded breccia which has undergone rheomorphic flow.

Some fragments within the breccia are composite. These range from a few millimetres to a few metres in diameter, and comprise breccia within breccia, fluidal-textured material in fluidal-textured material, glass within glass, or combinations of these. Some of the composite breccia fragments consist of fine breccia which resembles Sudbury Breccia.

Black Member

The Black Member (*see* Figure 15.10) is a laterally continuous unit of breccia 800 to 1200 m thick. Its lower contact with the Gray Member is generally gradational but locally sharp, and its upper contact with the Onwatin Formation is gradational. The Black Member is similar to the Gray Member in that the individual units of breccia have similar characteristics; the breccias are composed of similar types of fragments, including some with shock metamorphic features; the glasses and fluidal-textured material have similar overall characteristics; and the composite fragments have similar features.

The notable differences from the Gray Member are as follows.

1. The colour is due mainly to the presence of granular carbon of undetermined origin.
2. There is a subunit, at the base of most of the Black Member, rich in chloritized shards (Figure 15.12), which commonly contains very little carbon.
3. Various proportions of chloritized shards are present in much of this member.
4. Many of the glass fragments appear to be less altered and/or to have had different primary compositions (Muir and Peredery 1984).
5. Locally, in the lower third of the member, there are accretionary fragments.
6. Planar bedding is locally found in the upper third of the member and its orientation locally is at right angles to the regional strike.

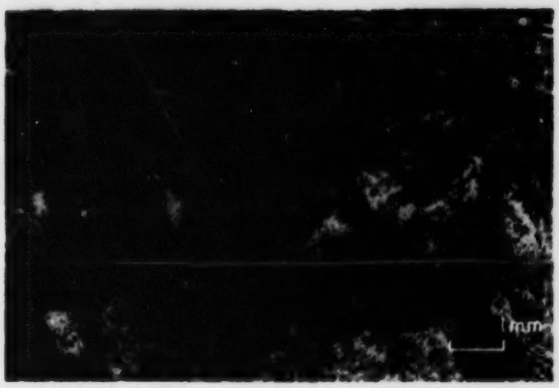


Figure 15.11. Gray Member. Finer parts of matrix contain many small, angular, felsic fragments. Most fragments are recrystallized felsic glasses.

7. The upper third of the member commonly has numerous rounded to subrounded blocks consisting of fine to medium sand-sized, recrystallized glass and rock particles.
8. The feldspars in the recrystallized glass and country rock fragments are mostly orthoclase and albite, whereas those in the Gray Member are generally microcline (upper third) and albite (lower two thirds) (Schandl et al. 1986).

The uppermost part of the Black Member, where it grades into the pelagic sediments of the Onwatin Formation, consists of redeposited, fine-grained, more or less stratified, locally disrupted Black Member material.

Melt Bodies

Melt Bodies generally form irregular masses, exhibiting gradational to sharp contacts with the Basal Member, and uniformly sharp contacts with the Gray and Black members, and the Sudbury Igneous Complex. The Melt Bodies locally occur as clusters within the Basal and Gray members, less commonly in the Black Member, and locally form protrusions into the Gray Member from lenses at the base of the Gray Member. The clusters range from 50 m to about 1 km long. Individual Melt Bodies range from a few metres to about 30 m across. Elsewhere, there are steeply dipping, flow-banded and/or spherulitic, aphanitic dikes in the Gray and Black members, apparently lacking any connection to a Melt Body.

Melt Bodies commonly have two parts, an igneous-textured, crystalline core and a chilled margin. The crystalline core is generally fine grained, massive, and rarely amygdaloidal. Locally, medium- to coarse-grained, acicular amphibole occurs either randomly oriented or as rosettes. Based on mineralogy and texture, the crystalline cores are either siliceous, feldspathic or granophyric (Peredery 1972b) and consist of various proportions of plagioclase, potassic feldspar, clinopyroxene, amphibole and chlorite after clinopyroxene, quartz, and accessory ilmenite, titanite or leucoxene, apatite and zircon. The aphanitic chilled margins of Melt Bodies may be massive, flow-banded, spherulitic or

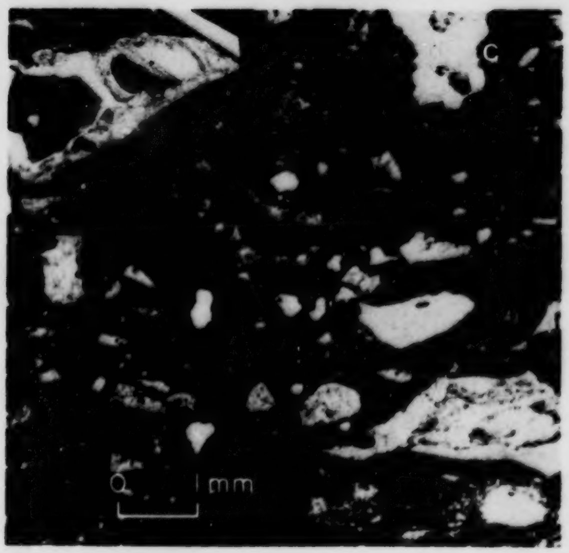


Figure 15.12. Black Member. Black matrix is rich in carbon particles. Chloritized shard-like fragments and (upper right corner at C) cryptocrystalline, felsic fragment.

brecciated, and range in thickness from several centimetres to a few metres. Country rock inclusions, within both cores and margins of Melt Bodies, account for less than 5% and up to 80% by volume. These generally occur as subrounded to subangular fragments of Huronian Supergroup quartz arenite and arkose although inclusions of other country rocks may also be present. Overall, Melt Body inclusions range in size from several metres across to submicroscopic, although the inclusions are generally well sorted by size within any given Melt Body.

Based on field relationships, mineralogy, textures and isotopic results, Muir (Discussion in Muir and Peredery 1984), Brockmeyer and Deutsch (1989), and Stevenson (1963) interpreted the Melt Bodies to be directly and genetically related to the Sudbury Igneous Complex. Based on whole-rock and mineral chemistry, and spatial association with shock metamorphosed rocks, Peredery (1972a) and Peredery and Morrison (1984) interpreted Melt Bodies to be the product of impact melting of country rock, a view shared by Brockmeyer and Deutsch (1989). These latter authors, however, interpreted the Sudbury Igneous Complex, the Melt Bodies and the Basal Member collectively as an impact melt system.

ONWATIN FORMATION

The Onwatin Formation (Coleman 1905; Burrows and Rickaby 1930; Thomson 1957; Martin 1957; Sadler 1958; Beales and Lozej 1975; Rousell 1984a) consists largely of massive to laminated claystone, siltstone and minor wacke. Locally, in the southwestern part of the Sudbury Basin near the base of the formation, there is a distinct unit known as the Vermilion Member, which consists mostly of carbonate rock, cherty carbonate rock, chert breccia, limestone, dolostone and claystone. Estimated thicknesses of the formation range from 1100 m (Coleman 1905), to 900 m (Arengi 1977), to 600 m (Rousell 1984a).

The lower contact of the Onwatin Formation grades into the Onaping Formation and is arbitrarily and variously defined by different authors. The upper contact is gradational with the Chelmsford Formation and is considered to be at the base of the first thick wacke bed that marks a change from predominantly mudstones to predominantly wackes (Rousell 1984a).

The Vermilion Member (Martin 1957; Rousell 1984a) overlies massive, siliceous and carbonaceous claystone, and has an average thickness of about 43 m. It comprises, from base to top: locally pisolitic, fine- to coarse-grained, commonly pyritiferous and base-metal-bearing carbonate rock, up to 30 m thick; cherty carbonate rock about 6 m thick; about 3 m of chert breccia consisting of black chert fragments in a white, recrystallized chert matrix; and 6 m of interbedded claystone, limestone, and dolostone. The Vermilion and Errington copper-zinc-lead-silver deposits (Burrows and Rickaby 1930; Martin 1957; Thomson 1957; Rousell 1984b) occur within carbonate rock. The extent of the Vermilion Member is uncertain: diamond drilling elsewhere in the North and South ranges intersected argillaceous limestone and/or cherty limestone from 1 to 30 m thick (Arengi 1977).

The remainder of the Onwatin Formation comprises massive to laminated claystone, fine-grained siltstone, silty claystone, coarse-grained siltstone and minor wacke, all of which are dark grey to black, commonly carbonaceous and pyritic. Laminae are mostly less than 5 mm thick, and coarser grained, graded beds are up to 15 cm thick. The wacke of this formation is similar in composition to wacke of the Chelmsford Formation.

The rock types that make up the Onwatin formation commonly contain quartz, plagioclase, micas (muscovite + chlorite ± biotite), clay minerals and fine-grained matrix material, indicating low-grade metamorphism. The carbon, most abundant in the claystone and siltstone, is very fine and resembles fossil algal and fungal filaments (Arengi 1977). Locally, metamorphism has resulted in the concentration of carbon into anthraxolite veins.

The pelagic rocks of the Onwatin Formation were deposited in a restricted basin with anoxic bottom conditions (Rousell 1984a). They grade upwards into the coarser sedimentary rocks of the Chelmsford Formation, indicating a transition to a higher energy, and probably somewhat shallower, depositional environment.

CHELMSFORD FORMATION

General descriptions of the Chelmsford Formation have been given by Burrows and Rickaby (1930), Williams (1957), Cantin (1971), Cantin and Walker (1972) and Rousell (1972), and are summarized here. The Chelmsford Formation comprises about 850 m (preserved thickness) of mostly wacke and siltstone. The lower contact with the Onwatin Formation coincides with the change from mudstone- to wacke-dominated sediments.

Many units of the Chelmsford Formation display the "A", "D" and "C" divisions of the Bouma cycle (decreasing order of abundance), and represent a sequence of proximal turbidite deposits. The turbidite beds are up to 5.2 m thick and average 1.23 m (Rousell 1972). Sedimentary structures present in the Chelmsford Formation are: carbonate concretions; rip-up clasts and other clasts; channel fillings; current marks; ripples, climbing ripples and crossbedding; convolute laminations; and load structures. Possible biogenic structures, similar to those formed by bottom-dwelling worms in more recent sedimentary rocks, have been described by Rousell (1972). Paleocurrent studies indicate that the predominant flow direction was to the southwest, parallel to the long axis of the Sudbury Basin (Cantin and Walker 1972; Rousell 1972).

The rock types composing these turbidites consist of dark grey, poorly sorted, commonly graded wacke; very fine-grained sandstone, siltstone and claystone; and buff, fine- to very fine-grained sandstone. The constituent minerals in the wackes are quartz, plagioclase, microcline, chlorite, muscovite, biotite and accessory minerals. The matrix is either light-coloured, consisting of fine-grained quartz and feldspar crystals, or dark-coloured, consisting of carbonaceous material. Shock metamorphic features in quartz and feldspar have not been observed.

MINERALIZATION IN THE WHITEWATER GROUP

Rousell (1984b) provided a review of mineralization in the Whitewater Group and presented new data as well. Mineralization of 3 ages occurs in the Group (Rousell 1983):

1. sulphide fragments in the Onaping Formation derived from pre–Sudbury Event rocks
2. mineralization formed from post–Onaping Formation to pre–Chelmsford Formation time (e.g., at the Vermilion and Errington mines)
3. mineralized quartz veins and anthraxolite veins

These mineralizations do not appear to be related to the nickel-copper deposits associated with the Sudbury Igneous Complex. Past attempts to mine the Vermilion and Errington deposits, and a few of the vein deposits, have been unsuccessful.

Mineralization in the Onaping Formation is characterized by irregularly distributed fragments, and/or disseminations, patches and grains of pyrrhotite (most common), chalcopyrite, pentlandite, sphalerite, galena, marcasite and pyrite. Although the sulphides generally make up only approximately 1% by volume, small, local occurrences contain up to 10%. Desborough and Larson (1970) determined the nickel content in pyrrhotite associated with this sulphide mineralization and found that the percentage of nickel was less than in the pyrrhotite of the Main Mass of the Sudbury Igneous Complex. The nickel content of pyrrhotite in the Onaping Formation is much lower than in the pyrrhotite of the nickel-copper ores at the base of the Sudbury Igneous Complex (Rousell 1984b).

The second type of mineralization in the Whitewater Group occurs in the Vermilion Member of the Onwatin Formation. It consists of pyrite, sphalerite, galena, chalcopyrite, marcasite and pyrrhotite. The ore within the closed Vermilion and Errington mines occurs as 2 types: disseminated fine-grained ore within the carbonate rock; and massive pyrite ore, generally high in zinc and low in copper, which locally replaces argillite at the base of the carbonate rock. The mineralization is interpreted to be sedimentary-exhalative in origin (Rousell 1984a).

Quartz veins are associated with these ores and also occur in the unmineralized rocks nearby. Those outside the Vermilion–Errington ores are barren, whereas those within the ore contain coarse-grained sphalerite, galena and chalcopyrite.

Sulphide-bearing quartz and quartz-carbonate veins make up the third type of mineralization in the Whitewater Group. Mineralized quartz veins are locally present in the Onaping Formation, mainly in the Basal Member. These veins contain various combinations and abundances of sphalerite, chalcopyrite, galena and pyrite. Small amounts of low-grade, gold-pyrite-bearing quartz veins were reported within the Onaping Formation (Basal Member) at the Gordon Lake property and the Creighton gold mine in the South Range (Rousell 1984b).

Sulphide-bearing, quartz-carbonate veins occur locally in a metagabbro sill, which is at or near the contact between the Onwatin and Chelmsford formations, in northern Balfour Township. Here, pyrite, arsenopyrite and chalcopyrite are present with minor gold and silver (Rousell 1984b). Locally, the arsenopyrite forms massive veins. Anthraxolite veins are locally found within the Onwatin Formation in the southwestern part of the Sudbury Basin. In the early years of this century, anthraxolite was investigated as a potential fuel.

The Sudbury Igneous Complex

The Sudbury Igneous Complex was formerly known as the Nickel-Bearing Eruptive, Sudbury Nickel Irruptive or Nickel Irruptive. A voluminous literature exists, spanning a century of research on the petrography, petrology, geochemistry, structure, and origin of the Complex (Giblin 1984b). The nomenclature and subdivision into Main Mass (norites, quartz gabbro and granophyres) and Sublayer (Contact Sublayer and Offset Sublayer) used here to describe the Igneous Complex are the same as those in Pye et al. (1984).

HISTORICAL OVERVIEW

One hundred years ago Bell (1891) published the first geological map of the Sudbury area on which rock units of the Sudbury Igneous Complex are described as greenstone, diabase and diorite breccia. Bell did not recognize the continuity of the mafic rocks around the Sudbury Basin. This continuity was first suggested by Walker (1897). Williams (1891) and von Foullon (1892) classified the mafic rocks as norite, and in 1904, Barlow (1904, 1908) published 2 maps of the South Range showing a twofold subdivision of the Complex into norite and granophyre. Barlow introduced the term "Nickel-Bearing Eruptive" for the Main Mass of the Sudbury Igneous Complex and Coleman (1905) proposed the term "offset" for dikes extending from the norite into the footwall. On an Ontario Bureau of Mines map, Coleman (1905) presented the Sudbury Igneous Complex as continuous around the Sudbury Basin. Burrows and Rickaby (1930) remapped the Sudbury Basin area for the Ontario Department of Mines. More recent compilations were produced by Card (1969) and Dressler (1984b).

In plan view, the Sudbury Igneous Complex has an elliptical shape with a northeast-trending long axis of 60 km and a short axis of 27 km (*see* Figure 15.1). Its three-dimensional shape, however, is not well known. The Complex has been considered a sheet-like laccolith (Coleman 1905), a sill (Collins 1934, 1935, 1936, 1937; Collins and Cooke 1938a, 1938b; 1947a, 1947b), a ring dike (Knight 1917; Williams 1957) or a lopolith (Dietz 1964; Fairbairn et al. 1968). A funnel shaped complex was suggested by, amongst others, Wilson (1956) and Naldrett and Kullerud (1967b), whereas Hamilton (1960) stated that it was an extrusive body.

An origin of the Sudbury Igneous Complex through magmatic differentiation was proposed by many: Barlow (1904, 1906); Coleman et al. (1929); Walker (1935) and Collins (1934, 1935, 1936, 1937); and Collins and Cooke (1938a, 1938b, 1947a, 1947b). Phemister (1926), however, believed that the norite and granophyre were separate intrusions, the norite being older.

Naldrett and Kullerud (1965, 1967a, 1967b) subdivided the norite into several units. Stevenson and Colgrove (1968) subdivided the norite into 4 "members". Dressler (1987) suggested that the common granophyre is older than the norite since, adjacent to a major fault at the southeastern end of the Sudbury Igneous Complex, norite and Contact Sublayer rocks are undeformed, whereas the granophyre has a strong schistosity parallel to this fault. Naldrett and Hewins (1984), in general, interpreted the Main Mass of the Complex to be the result of fractional crystallization of a mafic magma strongly contaminated by assimilation of silica-rich country rock.

Dietz (1964) suggested that the intrusion of the Complex was triggered by the impact of a meteorite. Dressler et al. (1987) stated that the Nipissing gabbro and the Sudbury Igneous Complex possibly have a common source and that a meteorite impact probably initiated remelting of this source rock, resulting in the generation of the Sudbury magma. More recent workers (Faggart et al. 1985; Stöffler et al. 1989) using Nd isotopic studies and theoretical considerations relating to large-scale impact events, interpreted the entire Sudbury Igneous Complex as an impact melt. However, Naldrett et al. (1988) pointed out that the isotope signatures are still consistent with a bulk assimilation model.

Souch et al. (1969) described the major ore-bearing gabbroic to quartz dioritic phase of the Sudbury Igneous Complex as the Sublayer. Pattison (1979), in a detailed account on the Sublayer, included the "granite breccia" as "leucocratic breccia" (now named Footwall Breccia) in the Sublayer.

The most detailed petrological account on the Main Mass of the Sudbury Igneous Complex was published by Naldrett et al. (1970). They applied modern igneous textural terminology to the major units. Naldrett and Hewins (1984), Naldrett et al. (1984) and Grant and Bite (1984) gave accounts of the Complex which are still current.

The following summary provides a current overview of the major units of the Sudbury Igneous Complex.

MAIN MASS

Naldrett et al. (1970) subdivided the rocks of the North Range and South Range of the Sudbury Igneous Complex into several units as illustrated in Figure 15.13.

The Main Mass of the Sudbury Igneous Complex is subdivided into a Lower, Middle and an Upper Zone (Pye et al. 1984; Dressler 1984b). The Lower Zone consists of mafic and felsic norites of the North Range, the South Range Norite and the quartz-rich norite of the South Range. The quartz gabbro of the Middle Zone occurs all around the Complex as does the granophyre (formerly called micropegmatite) of the Upper Zone (Naldrett et al. 1970; Naldrett and Hewins 1984).

Rocks of the South Range have been affected by a strong metamorphic overprint (Card 1978b; Dressler 1984a). Fleet et al. (1987) established that a low- to medium-pressure prograde event, that climaxed at medium grade immediately to the south of the Sudbury Igneous Complex, postdates the emplacement of the Complex and is probably Penokean in age.

The large-scale subdivisions of the Sudbury Igneous Complex do not exhibit fine-scale igneous layering as do many other large igneous complexes such as the Bushveld and the Stillwater complexes. The Sudbury Igneous Complex, compared to other large, mafic, igneous intrusions, is also unique in being abnormally rich in SiO_2.

The South Range Norite is a medium- to coarse-grained, black rock consisting of cumulus plagioclase and hypersthene and intercumulus quartz, augite, magnetite and ilmenite (Naldrett and Hewins 1984). Primary hornblende mantles pyroxenes. Some amphibole is also metamorphic in origin (Fleet et al. 1987). A faint igneous lamination is defined by the orientation of plagioclase crystals.

The quartz-rich norite is finer grained than the South Range Norite, contains up to 20% quartz and up to 25% biotite, and exhibits no igneous lamination. The quartz + granophyre content increases towards the lower contact. Pyroxenes are commonly uralitized.

The contact between the South Range Norite and the quartz gabbro of the Middle Zone is gradational (Naldrett and Hewins 1984); with a decrease upwards in the content of hypersthene, and an increase in quartz, augite, magnetite, ilmenite and apatite. Quartz and granophyric intergrowth of feldspar and quartz increase upwards in the Middle Zone.

The granophyre (Upper Zone) of the South Range is commonly strongly deformed and altered, but where slightly deformed it is very similar to the North Range granophyre. The observation of strongly deformed granophyre close to slightly, or undeformed quartz gabbro and norites of the South Range in Falconbridge Township, amongst other field evidence, led Dressler (1987) to suggest that the granophyre is older than the rocks of the Middle and Lower zones of the Sudbury Igneous Complex. However, Naldrett and Hewins (1984) described the contact of the granophyre with the underlying quartz gabbro as gradational, and Cowan and Schwerdtner (1990) observed deformed South Range Norite at Garson Mine west of Falconbridge Township.

The mafic norite of the Lower Zone, at the base of the Sudbury Igneous Complex in the North Range, is characterized by 40 to 60% hypersthene which commonly occurs as a cumulus phase enclosed by plagioclase. Hypidiomorphic textures of plagioclase and hypersthene also occur, mainly in the lower and upper parts of the mafic norite.

The felsic norite is the main rock type of the Lower Zone of the North Range. The rock is medium to coarse grained, and contains plagioclase and hypersthene as cumulus phases. Intercumulus minerals are augite, biotite and quartz. Granophyric intergrowth of feldspar and quartz is common. The upper third of the felsic norite does not contain hypersthene; instead, augite forms the cumulus phase (Naldrett and Hewins 1984). The contact with the overlying quartz gabbro is drawn where magnetite and apatite become

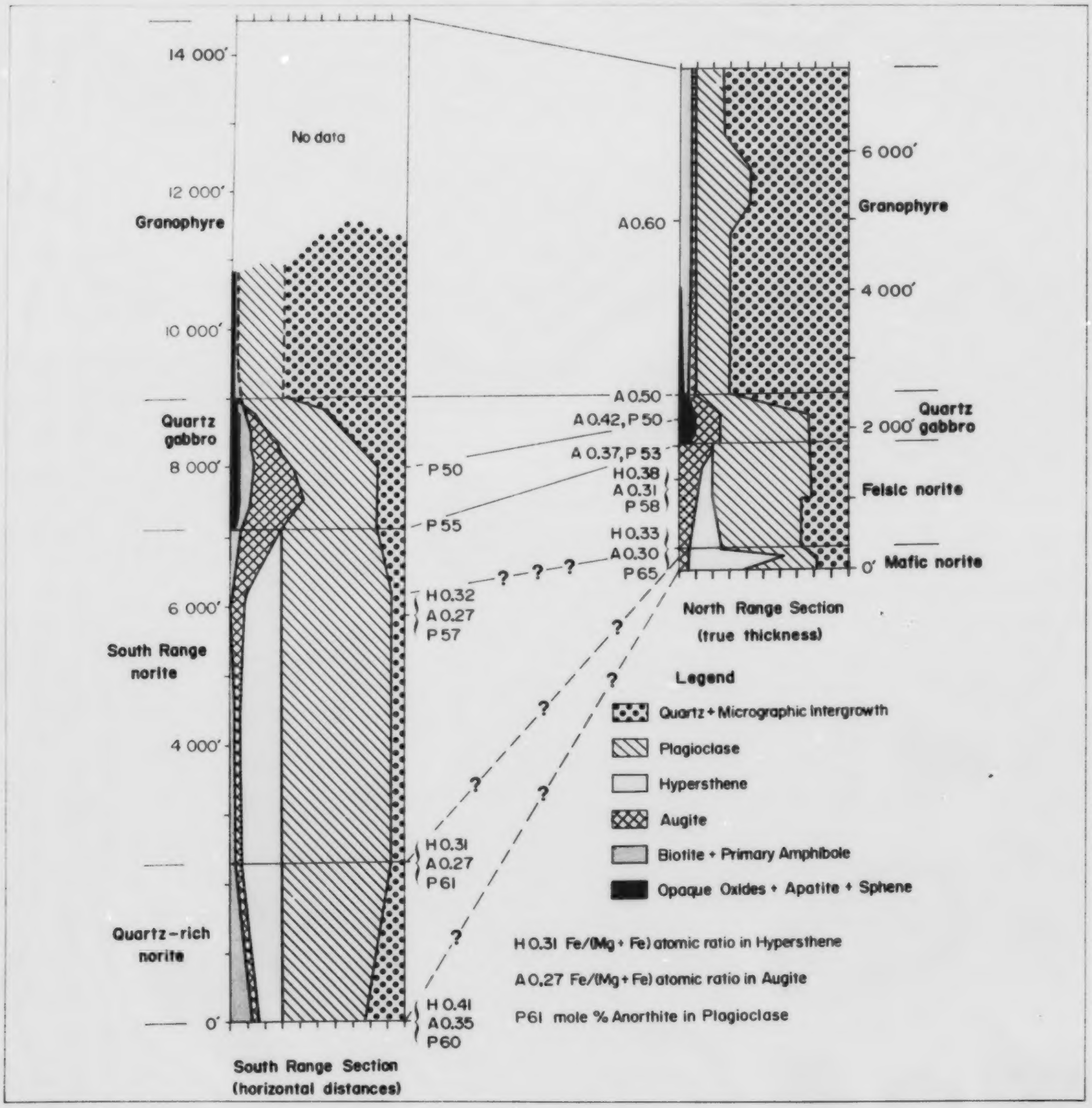

14 000'
12 000'
10 000'
8 000'
6 000'
4 000'
0'
No data
Granophyre
Quartz gabbro
South Range norite
Quartz-rich norite
P50
P55
H 0.32
A 0.27
P 57
H 0.31
A 0.27
P 61
H 0.41
A 0.35
P 60
South Range Section
(horizontal distances)
6 000'
4 000'
2 000'
0'
Granophyre
Quartz gabbro
Felsic norite
Mafic norite
A 0.60
A 0.50
A 0.42, P 50
A 0.37, P 53
H 0.38
A 0.31
P 58
H 0.33
A 0.30
P 65
?
North Range Section
(true thickness)
Legend
Quartz + Micrographic intergrowth
Plagioclase
Hypersthene
Augite
Biotite + Primary Amphibole
Opaque Oxides + Apatite + Sphene
H 0.31 Fe/(Mg + Fe) atomic ratio in Hypersthene
A 0.27 Fe/(Mg + Fe) atomic ratio in Augite
P 61 mole % Anorthite in Plagioclase

Figure 15.13. Diagram illustrating mineralogical variations on sections through the North and South ranges (*from* Naldrett et al. 1970).

common. Felsic norite also occurs in the South Range but it is rather insignificant, forming a zone only about 150 m thick below the base of the quartz gabbro.

The quartz gabbro of the North Range, Middle Zone is very similar to the quartz gabbro of the South Range. In the North Range, it contains more opaque oxide minerals, apatite and titanite (approximately 9%) than in the South Range (approximately 5%), and Naldrett and Hewins (1984) reported that exsolved ilmenite lamellae occur in the magnetite. In the quartz gabbro of the South Range, discrete ilmenite grains have been observed.

The granophyre, which makes up the Upper Zone of the Sudbury Igneous Complex, is commonly coarse grained and consists of plagioclase (about 1 part), granophyric intergrowth of quartz and feldspar (approximately 3 parts), and minor amounts of biotite and opaque oxide minerals. A plagioclase-rich granophyre phase containing about 35 modal % plagioclase has been described by Peredery and Naldrett (1975). It represents a more fractionated end member on the same mineralogical trends as those of the quartz gabbro. The more common, "normal" granophyre intruded below and into the plagioclase-rich granophyre.

Naldrett et al. (1970) and Naldrett and Hewins (1984) described in detail compositions of the rock-forming minerals of the Main Mass of the Sudbury Igneous Complex. Some of the results are presented in Figure 15.13.

Several gabbroic dikes occur within the Sudbury Basin where they intrude all formations of the Whitewater Group. They have not been observed intruding the Sudbury Igneous Complex. Most occurrences consist of strongly altered, carbonatized gabbros. Weakly altered gabbroic dikes are scarce; where present, they strongly resemble the felsic norite of the North Range or, where they contain inclusions of mafic rocks, they resemble the unmineralized rocks of the Sublayer (Dressler 1990). Geochemical investigations are under way to determine if these gabbro dikes within the Sudbury Basin are related to the Sudbury Igneous Complex.

Collins (1934) and Muir (1983) published the results of major element geochemistry of the Sudbury Igneous Complex. When compared with similar data from other large mafic igneous complexes all the rocks of the Sudbury Igneous Complex are abnormally rich in normative quartz in relation to the composition of plagioclases and pyroxenes. This observation led Irvine (1975) and Naldrett and Macdonald (1980) to suggest that the Sudbury Igneous Complex represented a magma strongly contaminated by silica-rich country rocks.

Kuo and Crocket (1979) studied rare earth element (REE) compositions of the Complex and showed that all igneous rocks of the Complex are enriched in light REEs . The $(La/Yb)_n$ ranges between 9 and 14. The REE data indicate that the rocks of the Sudbury Igneous Complex are related to each other by a fractional crystallization process, and that a silicic component with a high La/Yb ratio was involved in the generation of the magma.

Strontium isotopic geochemistry has been studied by many researchers, such as Fairbairn et al. (1960, 1965, 1968), Souch et al. (1969), and Hurst and Wetherill (1974). A detailed study by Gibbins and McNutt (1975) defined isochron ages of 1.956±0.098 Ga (initial $^{87}Sr/^{86}Sr$ of 0.7063±0.004) for the norite and 1.680±0.062 Ga (initial $^{87}Sr/^{86}Sr$ of 0.7083±0.0014) for the granophyre. These ratios are relatively high for mafic rocks of this age and may indicate magma contamination.

Uranium-lead age determinations on zircon and baddeleyite from North Range felsic and mafic norites, the South Range Norite and the granophyre indicate that the Main Mass of the Sudbury Igneous Complex has an age of 1850±1 Ma (Krogh et al. 1984).

SUBLAYER

The Sublayer is a mafic to intermediate rock occurring at the base of the Sudbury Igneous Complex (Contact Sublayer) and host to much of the nickel-copper sulphide ores of the Sudbury Structure (Yates 1948). It also occupies dikes in the country rocks of the Sudbury Structure (Offset Sublayer) which are termed "offset dikes." (Figure 15.14). Naldrett and Kullerud (1967a, 1967b) termed the unit "xenolithic norite", Souch et al. (1969) called it "igneous sublayer." Pattison (1979) described the Sublayer and included in it the barren and ore-bearing "leucocratic breccias" (i.e., Footwall Breccia). These breccias occur at the contact of the Sudbury Igneous Complex with the footwall rocks at several locations in the North and East Ranges. The Contact Sublayer has been described in detail by Naldrett et al. (1984). The offset dikes have been described by Grant and Bite (1984).

Contact Sublayer

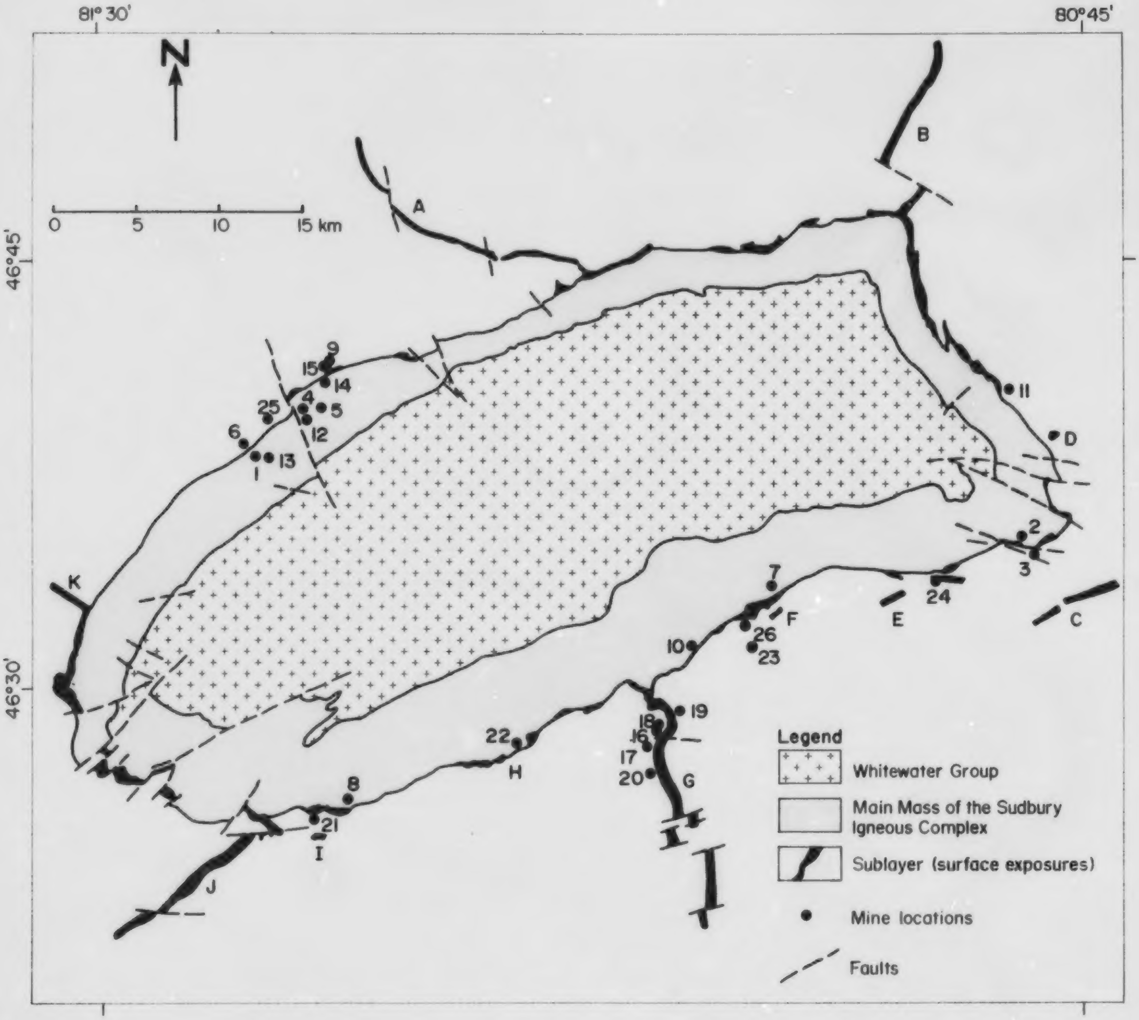
The Contact Sublayer is a fine- to medium-grained quartz dioritic to noritic unit forming discontinuous, sheet-like bodies, more or less parallel to the lower contact of the Sudbury Igneous Complex, and protrusions and embayments into the footwall rocks at the base of the Complex. It is in sharp contact with the overlying norites and in sharp to gradational contact with the Footwall Breccia which locally underlies the Sublayer.

The gabbroic-quartz dioritic matrix of the Contact Sublayer phases commonly is fine to medium grained and consists of hypersthene and augite in a ratio of 2:1 to 1:2 making up 40 to 60% of the rock. Lath-shaped plagioclase, minor quartz and potassic feldspar make up the remainder of the matrix. Inclusions constitute 5 to 80% of the Contact Sublayer.

There is field evidence that indicates that the Sublayer consists of at least 3 subunits, defined by xenolith type and intrusive relationships (Dressler 1982; Naldrett et al. 1984). The oldest subunit is the main, sulphide-bearing, Sublayer subunit. The younger subunits contain little or no sulphide mineralization (Figure 15.15).

The sulphide-bearing, oldest Sublayer subunit is characterized by subrounded to rounded, mafic, ultramafic, sulphide and, in places, country rock inclusions (Figure 15.16). Sulphides also form disseminations or a massive matrix to the igneous xenoliths. The age of this ore-rich subunit with respect to the norites of the Main Mass of the Sudbury

Igneous Complex is not known with certainty, but Naldrett and Kullerud (1967b) and Naldrett et al. (1972) stated tentatively that the Sublayer is younger than the norite, an observation supported by Yates (1948). The younger subunits of the Sublayer (Dressler 1982; Naldrett et al. 1984) are characterized by inclusions of footwall rocks and, in places, contain xenoliths derived from the Lower Zone of the Sudbury Igneous Complex and the older, sulphide-rich Sublayer.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Legend
Whitewater Group
Main Mass of the Sudbury Igneous Complex
Sublayer (surface exposures)
Mine locations
Faults

Mines of Falconbridge Limited

1 Boundary Mine
2 Falconbridge Mine
3 Falconbridge East Mine
4 Fecunis Mine
5 Fraser Mine
6 Hardy Mine
7 Lindsley property
8 Lockerby Mine
9 Longvack South Mine
10 McKim Mine
11 Nickel Rim Mine
12 North Mine
13 Onaping Mine
14 Strathcona Mine

Mines of Inco Limited

15 Coleman Mine
16 Copper Cliff Mine
17 Copper Cliff No. 1 Mine
18 Copper Cliff No. 2 Mine
19 Copper Cliff North Mine
20 Copper Cliff South Mine
21 Crean Hill Mine
22 Frood-Stobie Mine
23 Garson Mine
24 Levack Mine
25 Little Stobie Mine
26 McCreedy West Mine

Offset Dikes

A Foy
B Whistle
C Manchester
D Maclennan
E Kirkwood and McConnell
F Frood-Stobie
G Copper Cliff
H Creighton Embayment
I Vermilion
J Worthington
K Ministic

Figure 15.14. Distribution of the Sublayer of the Sudbury Igneous Complex and approximate location of major mines.

Figure 15.15. Sublayer (younger subunit) of Sudbury Igneous Complex. Clasts are locally derived igneous and supracrustal rocks in a quartz diorite matrix. Near Whistle property, northeastern corner of Sudbury Igneous Complex.

Figure 15.16. Sublayer (older subunit) of Sudbury Igneous Complex. Clasts are mafic igneous rocks of unknown derivation. They are set in a quartz diorite matrix. The matrix contains fine-grained, disseminated sulphide minerals.

Therefore, these subunits are "post-norite" in age. They contain no ultramafic fragments. Fifty to ninety percent of these inclusions are fine-grained metavolcanic rocks, possibly derived from the lower Huronian Supergroup. The remainder reflect rock types found in the nearby footwall (Naldrett et al. 1984; Lafleur and Dressler 1985).

Offset Dikes

The offset dikes of the Offset Sublayer either extend approximately radially from the Sudbury Igneous Complex or lie concentrically in the country rocks. Several of the dikes, amongst them the Frood–Stobie and Manchester offsets, were injected into pre-existing dike-like bodies of Sudbury Breccia. Quartz diorite is the dominant rock type and, as with the Contact Sublayer, is characterized by a variety of inclusions and sulphide mineralization.

Grant and Bite (1984) provided a comprehensive summary on the offset dikes and much of the following is based on their work. They subdivide the quartz diorite offsets into 3 types, according to their orientation with respect to the Sudbury Igneous Complex (*see* Figure 15.14). The first type, the radial offsets, commonly begin at embayments of Contact Sublayer and extend into the surrounding footwall rocks for several kilometres, like the spokes of a wheel. These radial dikes are the Copper Cliff, Whistle, Foy, Ministic and Worthington offsets. The second type, concentric dikes, for example the Frood–Stobie and Manchester offsets, strike more or less parallel to the contact of the Sudbury Igneous Complex. Discontinuous offset dikes, the third offset type described by Grant and Bite (1984), occur in Maclennan Township and at Creighton. All dikes dip steeply to vertically, and have sharp contacts with the host rocks.

Three types of quartz diorite occur in the dikes:

1. The hypersthene quartz diorite is medium to coarse grained, and consists of acicular hypersthene, plagioclase laths, with interstitial quartz, potassium feldspar, and granophyric intergrowth of feldspar and quartz. Biotite, apatite, titanite, ilmenite, and leucoxene are accessory minerals. Hypersthene commonly has hornblende rims.
2. The two-pyroxene quartz diorite is finer grained than the hypersthene quartz diorite and contains blocky clinopyroxene in addition to hypersthene (Grant and Bite 1984).
3. The amphibole-biotite quartz diorite is characterized by amphibole and biotite pseudomorphs after pyroxene, suggesting that it is an alteration product of the 2 other quartz diorite types. This third type of quartz diorite is the most common, and the principal one in the radial dikes.

All 3 types of quartz diorite contain inclusions similar to those in the Contact Sublayer, and several phases of Sublayer have been observed in one and the same offset dike, similar to the subunits in the Contact Sublayer. Inclusions of Footwall Breccia and norite derived from the Main Mass of the Sudbury Igneous Complex have also been observed in this environment. Footwall Breccia, in places, appears to fill the Foy Offset from margin to margin (Pattison 1979), suggesting either that quartz diorite magma intruded along zones of weakness previously filled by the older Footwall

Breccia, or that Footwall Breccia material was transported to its present position by the quartz diorite magma. Sulphide minerals appear to be spatially, and possibly also genetically, associated with mafic and ultramafic exotic inclusions. This association and the intermediate composition of the quartz diorite suggest that the quartz diorite was not the source for the sulphides. The mineralization was passively transported to its present position by the quartz diorite magma. (Grant and Bite 1984).

The Sudbury Orebodies

The Sudbury orebodies are unique and constitute the largest known concentration of nickel-copper sulphides on Earth. They occur along the lower contact of the Sudbury Igneous Complex, that is, in the Contact Sublayer, the Footwall Breccia and/or the footwall rocks themselves, and in the offset dikes. Short summaries on the various deposit types and their mineralogy, composition and genesis are presented here. Again, much of the information has been taken from papers in Pye et al. (1984)

DEPOSIT TYPES

The most recent summary on the ore deposits of the Sudbury Structure, excluding those of the Whitewater Group, is by Naldrett (1984a) who subdivided the deposits into 5 groups: 1) South Range deposits, 2) North Range deposits, 3) offset deposits, 4) fault-related deposits, and 5) miscellaneous deposits.

South Range Deposits

The Murray Mine (Naldrett 1984a, p.303 and Figure 1) and the Little Stobie Mine (Davis 1984) are typical examples of South Range deposits. In the Murray Mine, massive inclusion-bearing sulphide is overlain by Contact Sublayer quartz diorite which contains up to 60% disseminated sulphide. The inclusions are exotic mafic and ultramafic rocks and footwall rocks. The sulphide content decreases upward in the quartz diorite to about 20%. The deposit straddles the contact between the Sudbury Igneous Complex and the footwall rocks, which are supracrustal rocks of the Huronian Supergroup and granite of the Murray Pluton. Contacts between the deposit and the footwall rocks and the norite of the Main Mass of the Complex are well defined and sharp.

North Range Deposits

The North Range deposits occur mainly in the Onaping–Levack area (Coats and Snajdr 1984). Several of them are very rich. The ore deposits, according to Coats and Snajdr (1984), are spatially related to embayments in the footwall; which are filled by massive sulphides, mineralized Footwall Breccia, Sublayer rocks and mafic norite. Ore also occurs within the footwall rocks proper, either as large massive bodies, or as dikes and vein networks. Chalcopyrite-rich dikes and vein networks constitute the "copper zones" (Abel et al. 1979). Economically, Footwall Breccia is the most important host for the North Range mineralizations. Sulphides in the breccia occur as stringers, disseminations and blebs. Massive ore bodies within the Footwall Breccia, in places, are brecciated and may contain breccia and country rock inclusions.

Offset Deposits

The deposits of the Copper Cliff Offset (Cochrane 1984) are examples of the "offset deposit" type. The deposits in this offset are either at or near its centre, or at the contact of the dike with the country rocks. The economic sulphide concentrations are spatially associated with inclusion-rich quartz diorite and with local thinning of the dike. Sulphides form blebs in the quartz diorite matrix. Cochrane (1984) described a zoning in ore composition characterized by "an increase in chalcopyrite content with increasing distance from discontinuities in the dike." He interpreted this zoning to reflect fractional crystallization of a sulphide melt.

Fault-Related Deposits

The East, Falconbridge and Garson mines belong to the so-called "fault-related deposits." Two of these mines, the Falconbridge and East mines, are located at the southeastern margin of the Sudbury Igneous Complex (Owen and Coats 1984). The deposits are within fault zones at the contact of the Complex with metavolcanic and metasedimentary rocks of the Stobie Formation of the Huronian Supergroup. The ore zones consist of 2 types. The "Main Zone contorted schist inclusion sulphide" (Owen and Coats 1984), is a sulphide breccia containing inclusions of norite (from the Main Mass of the Sudbury Igneous Complex), and Huronian supracrustal rocks. The main ore minerals are pyrrhotite, pentlandite and chalcopyrite. The second ore type, the "Southwall inclusion massive sulphide zones" (Owen and Coats 1984), lies to the south, near or at the contact with the "Main Zone" deposits, and contains inclusions of Huronian supracrustal rocks, quartz and jasperoid. Silicification of the footwall rocks, strong deformation of much of the mineralization, intense alteration of the Contact Sublayer, and late fractures cutting across the deposits filled with sphalerite, marcasite, galena and carbonates, are all indications of tectonism and hydrothermal activity postdating and possibly, in part, predating the emplacement of the deposits (Owen and Coats 1984). The primary magmatic source of the mineralization is considered to be the Contact Sublayer (Owen and Coats 1984).

Miscellaneous Deposits

The fifth group of Sudbury ore deposits, the "miscellaneous group", is not well documented in the literature. In it, deposits are grouped together that are mainly characterized by postemplacement modifications. One such deposit is the McKim Mine described by Clark and Potapoff (1959). Part of the McKim mineralization is enclosed in brecciated Murray granite, which is considerably older (2223 Ma, Krogh et al. 1984) than the Sudbury Igneous Complex (1850 Ma, Krogh et al. 1984). Near the mine, granitic veins from the Murray Pluton cut into the Complex. These veins are products of remobilization caused by the heat of emplacement of the Sudbury Igneous Complex magma. The remobilization and brecciation of the granite has apparently allowed sulphide fluids from the Sublayer to migrate into the granite. The mineralized zones consist predominantly of pentlandite and pyrite in the granite, with quartz and carbonate as gangue minerals.

MINERALOGY AND COMPOSITION OF THE DEPOSITS

The most common minerals in the Sudbury ores are pyrrhotite, pentlandite, chalcopyrite, pyrite and titanium-poor magnetite. A large number of other minor ore minerals are known from Sudbury (Table 15.2, Naldrett 1984b). Coats and Snajdr (1984) published a list of minerals identified in the copper zones of Strathcona Mine. Several typical Sudbury ore samples are shown in Figure 15.17.

Many Sudbury deposits are characterized by compositional zoning (Figure 15.18), and there are marked compositional differences between various Sudbury orebodies. Fractional crystallization of monosulphide solid solution from a sulphide melt (Naldrett 1984b) may account for the compositional zoning observed. The compositional differences between various Sudbury deposits may be the result of incomplete preservation of the orebodies (Naldrett 1984b). In the North Range deposits, for example at Strathcona Mine, fractional crystallization of a sulphide melt might explain the copper, platinum, palladium and gold enrichment in the ores at greater distances from the source Sublayer magma. Ores enriched in these elements are absent at Falconbridge Mine in the South Range. Their absence possibly can be explained by post-ore tectonic removal (Owen and Coats 1984).

GENESIS OF THE SULPHIDE DEPOSITS

The discussions concerning the genesis of the ores have been ongoing for the last century . Various ideas have been presented. Differentiation of the ores from a mafic magma was proposed by Bell (1891), Barlow (1904, 1906) and Thode et al. (1962). A hydrothermal origin was put forward by Dickson (1903), Knight (1917), Wandke and Hoffman (1924), Davidson (1948) and Lochhead (1955). Howe (1914) and Bateman (1917) suggested that the ores were injected as sulphide melts. Naldrett and Kullerud (1967b) suggested that the sulphur of the ores was derived from external sources. Kullerud and Yoder (1963) saw immiscibility of quartz diorite and sulphide liquids as having played a major role in the formation of the Sudbury orebodies, a view, in general, also forwarded by Hawley (1962). Souch et al. (1969) showed that the mineralization occurred in a marginal facies of the Sudbury Igneous Complex, the Contact Sublayer and in the offset dikes. They concluded that the ores were products of magmatic crystallization of a sulphur-rich, mafic magma.

Naldrett (1984a) stated that many geologists have been persuaded "that the ores are indeed magmatic in origin." This is based on the basal nature of the ore zones within the Sudbury Igneous Complex; the concentration of cobalt and platinum group elements in addition to nickel and copper; the absence of lead and zinc, which are common in some hydrothermal alteration; and "the presence of textures highly suggestive of the coexistence of a magmatic sulphide liquid and partially crystalline silicate magma" (Naldrett 1984a).

Irvine (1975) suggested that the assimilation of silica-rich material by a mafic magma would lower the solubility of sulphur within it, and proposed that this had occurred within the Sudbury Igneous Complex. Naldrett and Macdonald (1980) pointed out that the silica-rich nature of the Complex supports this hypothesis. Naldrett et al. (1988), based on REE, Nd and Sr isotope investigations, and major and trace element modelling, stated that the mafic Sudbury parent magma has been contaminated by the assimilation of 33 to 75% of silica-rich country rocks. Strong brecciation, pulverization and shock melting of the country rocks prior to the emplacement of the Complex would have greatly facilitated the assimilation of such a large amount of material.

STRUCTURAL GEOLOGY

The Sudbury Structure lies within the southern Canadian Shield and has been strongly affected by the Penokean Orogeny. Mineral and whole-rock radiometric ages obtained from Proterozoic rocks of the Sudbury area span a long period of time from approximately 2450 Ma to about 1200 Ma (Stockwell 1982; Krogh et al. 1984). An unpublished, radiometric whole-rock Rb-Sr age (1950 Ma) from the Archean Levack gneisses near the Sudbury Igneous Complex in Levack Township indicates that the rocks north of the Igneous Complex have also been affected by the Penokean Orogeny (B.V. Rao, University of Toronto, personal communication, 1984).

These orogenic events have been described elsewhere in this volume (Bennett et al., this volume). This chapter deals only with their structural effects on the Sudbury Structure. These effects were confined to the South Range and central part of the Sudbury Basin. Understanding the shape of the Sudbury Basin and the Sudbury Igneous Complex before deformation is important, as it has some relevancy to the origin of the Sudbury Structure (Tables 15.3 and 15.4).

The Sudbury Structure acquired its present elliptical shape during northwesterly directed thrusting (Shanks 1991). This thrusting event was responsible for several strike-parallel faults (i.e., faults parallel to the strike of rock units within the central Sudbury Basin) cutting across the Sudbury Igneous Complex and the rocks of the Whitewater Group, from the southwest to the southeast (Thomson 1957; Card 1968; Dressler 1984b). The thrusting event was also responsible for a major reverse shear zone, the Fairbank–Whitson Lakes Zone (Shanks 1991) (Figure 15.19). This zone was probably produced in the same regional field of stress that was responsible for thrusts and reverse faults in the rocks of the Huronian Supergroup south of the Sudbury Igneous Complex (Shanks 1990). An apparent sinistral strike separation of about 3.2 km occurs along the Cameron Creek Fault, one of the major strike-parallel reverse faults at the southwestern end of the Sudbury Igneous Complex (*see* Figure 15.19). An apparent dextral separation of almost 1 km occurs along an unnamed reverse fault just north of the Cameron Creek Fault. These faults dip southeasterly, but the

Table 15.2. Minerals of Sudbury ores (*from* Naldrett 1984b)

MINERAL SPECIES	FORMULA	REFERENCES	NOTES
MAJOR			
Pyrrhotite			
hexagonal monoclinic	$Fe_{(1-x)}S$ $Fe_{(1-x)}S^{-1}$	Cowan (1968) Corlett (1971)	Distribution of hexagonal and monoclinic forms studied intensively at Strathcona Mine
Pentlandite	$(Fe,Ni)_9S_8$	Hawley and Stanton (1962)	Ni content 33 to 35 wt %; Co content 1 wt %
Chalcopyrite	$(Cu,Fe)S_2$	Hawley and Stanton (1962)	
Magnetite	Fe_3O_4	Hawley and Stanton (1962) Naldrett and Kullerud (1967b)	Magnetite in the ores is low in TiO_2
Pyrite	FeS_2		Types: 1 Pink Nickelean 2 Early Resorbed 3 Exsolved 4 Hypogene Replacement
MINOR			
Cubanite	$CuFe_2S_3$	Hawley and Stanton (1962) Abel et al. (1979)	Particularly in the Frood–Stobie deposit and in copper zone at Strathcona
Galena	PbS	Hawley and Stanton (1962)	As small veins in fractures and as small flecks in pyrite
Sphalerite	ZnS	Hawley and Stanton (1962)	With chalcopyrite as late veins and fault filling
Millerite	ZnS		Originally regarded as secondary. Hypogene mineral found in footwall at Strathcona Mine
Niccolite	NiAs	Hawley and Stanton (1962)	At Worthington, Frood and Garson in association with other aresenides
Gersdorffite	NiAsS	Hawley and Stanton (1962) Cabri and Laflamme (1976)	With chalcopyrite and as late-stage veins
Cobaltite	CoAsS		Rare
Gold	Au	Hawley and Stanton (1962) Michener (1940)	Found in siliceous mineral zone at Frood Mine
Silver	Ag	Hawley and Stanton (1962)	Near base of Frood Mine
Bismuth	Bi	Hawley and Stanton (1962)	Associated with galena or with parkerite and bornite at Frood Mine
Tetradymite	Bi_2Te_3S	Hawley and Stanton (1962)	Commonly associated with veins of galena
Hessite	Ag_2Te	Hawley and Stanton (1962)	At Frood Mine
Maucherite	$Ni_{11}As_8$	Hawley and Stanton (1962)	Commonly occurs with niccolite and gersdorffite
Bornite	Cu_5FeS_2	Hawley and Stanton (1962)	Only observed in copper-rich zones at Frood deposit
Parkerite	Ni_3Bi_2S	Michener (1940)	
Schapbachite	$AgBiS_2$	Hawley and Stanton (1962)	Frood Mine—1 grain observed
Altaiite	PbTe	Cabri and Laflamme (1974, 1976)	Found at Coleman, Frood and Crean Hill mines, commonly associated with PGM
Argentiferous pentlandite	$(Fe,Ni)_8AgS_8$	Cabri and Laflamme (1976)	Light pink, probably confused with bornite
Breithauptite	$Ni_7(Bi,Sb,Te)_2S_8$		
Hauchecornite			
Mackinawite			

Table 15.2. Minerals of Sudbury ores (*from* Naldrett 1984b)

MINERAL SPECIES	FORMULA	REFERENCES	NOTES
Wehrlite	$(BiPb)_{105-120}$ $(TeSb)_{0.95-0.72}$	Cabri and Laflamme (1976)	Associated with chalcopyrite, cubanite
PLATINUM GROUP MINERALS (in order of abundance)			
Michenerite	PdBiTe	Cabri and Laflamme (1976) Hawley and Berry (1958)	Principal Pd mineral at Sudbury and occurs in most deposits
Moncheite	$PtTe_2$	Cabri and Laflamme (1976)	Observed at Creighton and Levack West
Sperrylite	$PtAs_2$	Hawley and Stanton (1962) Cabri and Laflamme (1976)	Most common Pt mineral of South Range
Insizwaite	$PtBi_2$	Cabri and Laflamme (1976)	At Coleman Mine
Sudburite	PdSb		At Copper Cliff South and Frood mines
Froodite	$PdBi_2$	Hawley and Berry (1958) Cabri and Laflamme (1976)	Frood, Vermilion, Creighton, Levack West and Coleman mines
Kotulskite	PdTe	Cabri and Laflamme (1976)	Levack West and Creighton mines
Niggliite	PtSn	Cabri and Harris (19762)	Coleman Mine
Merenskeyite	$PdTe_2$	Cabri and Laflamme (1976)	Creighton, Crean Hill and Levack West mines
Mertieite II	$Pd_8(SbAs)_3$	Cabri and Laflamme (1976)	Creighton Mine
Unnamed	Ps(Bi,Sb,Te)	Cabri (1973) Cabri and Laflamme (1976)	Vermilion Mine
Unnamed	$Ag_4Pd_3Te_4$	Cabri and Laflamme (1976)	Levack Mine West
Palladian melonite	$(Ni,Pd)(Te,Bi)_2$	Cabri and Laflamme (1976)	Falconbridge, Strathcona, Crean Hill and Creighton mines
SECONDARY			
Violarite	$FeNi_2S_4$	Hawley and Stanton (1962)	In many ores subject to weathering
Marcasite	FeS_2	Hawley and Stanton (1962)	Replacement bodies in crosscutting seams and fractures
Valleriite	$4(Fe,Cu)\cdot 3(Mg,Al)\cdot (OH)_2$	Hawley and Stanton (1962)	Found in chalcopyrite–rich samples

Table 15.3. Factors supporting meteorite impact origin of the Sudbury Structure (B.O. Dressler).

1. Very strong brecciation of country rocks occurs up to 80 km from the Sudbury Igneous Complex; such features are not known to be associated with any accepted endogenic structure.
2. Shock metamorphic features are present in basement rocks, indicating shock pressures of greater than 20 GPa. These features are planar elements ("shock lamellae") in quartz and feldspar (not found in undisputed endogenic structures, French 1990). Planar elements are also known from mineral and rock fragments of the Onaping Formation. Recrystallized glass fragments in this formation are interpreted as diaplectic glasses, glasses formed by shock and not by heat.
3. Shatter cones comparable to those at Sudbury are present in many undisputed impact structures.
4. Collar rocks of South Range structure are overturned, as observed in some impact structures. (Shoemaker 1963; Offield and Pohn 1977; Daly 1947).
5. Most undeformed impact structures on Earth, and on other planets and moons of the solar system, are circular or subcircular. A small number of noncircular, interpreted impact structures are known on the moon (e.g., Schiller and Messier structures). They are believed to have been caused by very low angle impact or by the impact of more than one projectile. If this interpretation is correct, noncircularity does not negate an impact origin for the Sudbury Structure. Furthermore, recent structural investigations by Shanks (1991) suggest that the Sudbury Structure possibly was circular before deformation.

Table 15.4. Factors supporting an endogenic origin of the Sudbury Structure (T.L. Muir).

1. Impact processes are sudden and short-lived. Evidence in the Sudbury Structure for multiple explosive and depositional events over an extended time are as follows: megascopic and mesoscopic stratification in the upper members of the Onaping Formation; composite fragments of breccia ("breccia within breccia") and recrystallized fluidal glasses ("glass within glass") in the upper members of the Onaping Formation; Melt Body dikes and steeply dipping, tabular dikes of fluidal, recrystallized glasses intrude consolidated breccias of the upper members of the Onaping Formation (Muir 1984) and; Sudbury Breccia dikes cutting across other Sudbury Breccia dikes, and fragments of Sudbury Breccia within Sudbury Breccia (Muir 1983).
2. Melt Bodies form flows and intrusive bodies within the upper members of the Onaping Formation (Peredery 1972b). Igneous-textured melt bodies are similar to a phase of granophyre (texturally, mineralogically and/or chemically), considered to be genetically related to the Sudbury Igneous Complex (Stevenson 1963; Muir 1984) which is widely considered to represent endogenic magmatism.
3. Geochemical zoning of the matrix of Onaping Formation breccias, from dacitic-andesitic in the Gray Member to basaltic in the Black Member; and the presence of euhedral to subhedral, phenocryst-like crystals; as well as fragments of recrystallized glasses of mixed compositions, in both members; could represent eruptive products of a zoned magma chamber and/or mixed magmas (Muir 1984).
4. The Sudbury Structure is considered to be an integral part of the regional geology (Card and Hutchinson 1972; Card et al. 1984; Muir 1984). Using accepted geological field relationships, various endogenic events that collectively predate and postdate the "Sudbury Event" can account for several major features such as outliers of Huronian rocks (crudely concentric to the northern margin of the Sudbury Igneous Complex), high-grade gneisses in the footwall rocks of the North Range, and locally overturned beds in the South Range footwall rocks (*see* Card et al. 1984; Muir 1984).

angle of dip is not known. Rousell (1984c) estimated the dip separation on the Cameron Creek Fault as roughly 2800 m and the throw near 2600 m. At the southeastern corner of the Sudbury Basin, on the Norduna Fault (*see* Figure 15.19), an approximately 4 km wide northern section of the Sudbury Igneous Complex is in contact with a 7.5 km wide southern section. Here, a thicker and originally deeper, southern part of the Sudbury Igneous Complex abuts against a thinner, northern and upper portion of the Complex. The dip of the reverse Norduna Fault is not known. The Airport Fault (*see* Figure 15.19), north of the Norduna Fault, is a sinistral fault which, surprisingly, dips 70° to the northeast (P. Snajdr, Falconbridge Limited, personal communication, 1990). The Bailey Corners Fault (*see* Figure 15.19), south of the Norduna Fault, dips approximately 65° to the southwest (Norduna North Project Plan, courtesy of Falconbridge Limited). The No.1 and No.2 faults just south of the Bailey Corners Fault (not shown on Figure 15.19, *see* Dressler 1984b) dip 45° to the northwest and 30° to the southwest respectively (P. Snajdr, Falconbridge Limited, personal communication, 1990).

In general, the magnitude of net displacement on the large intrabasin strike-parallel faults is not known. Also unknown are the structural conditions at both the eastern and western ends of the strike faults. These faults may be branches of the regional Murray Fault System (Card and Hutchinson 1972), or may be cut by this system. They probably had some effect on the present shape of the Sudbury Structure. How great this effect really was, is unknown, since the original shape of the Sudbury Igneous Complex and the dip separations along the faults are in dispute (*see* below).

The northwesterly striking faults of the Onaping fault system (Card and Hutchinson 1972) that cut across the North Range and the north-central part of the Sudbury Structure had little or no effect on the overall shape of the Structure. The 2 most prominent of these faults are the Fecunis Lake and the Sandcherry faults (for location *see* Figure 15.19). They strike north-northwesterly and have an apparent sinistral strike separation at the Sudbury Igneous Complex of approximately 1 km and 0.7 km respectively. Northwesterly trending lineaments occur in the Sudbury Igneous Complex in the South Range, and may belong to the Onaping fault system.

Several attempts have been made to calculate the horizontal shortening of the northwest-southeast axis of the Sudbury Structure (Peredery 1972b; Rousell 1972; Brocoum and Dalziel 1974; Clendenen 1986) using structures such as carbonate concretions in the Chelmsford Formation. Estimates of total shortening range from 50% (Brocoum and Dalziel 1974) and 40% (Clendenen 1986) to 36% layer-shortening strain (Shanks 1991). Folding within the Chelmsford Formation caused approximately 10% shortening of the central part of the Sudbury Basin (Shanks 1991). On northwest-southeast vertical sections across the Fairbanks–Whitson Lakes Zone, Shanks (1991) estimated the horizontal shortening caused by continuous ductile shear to be between 6 km and 13 km. The slaty sediments of the Onwatin Formation may have accommodated 30 to 80% shortening.

Based on Shanks' (1991) information described above, and his estimate of 50% northwest-southeast shortening of the Sudbury Igneous Complex and Sudbury Basin, the dimensions of the Structure before shortening were approximately equivalent; that is, the Sudbury Structure was circular before deformation. This opinion is shared by Brocoum and Dalziel (1974), but is in conflict with studies by Cantin and Walker (1972), Rousell (1972) and Muir (1984), who maintain that the Structure was never circular. Cantin and Walker (1972) and Rousell (1972) based their opinion on paleocurrent directions in the rocks of the Chelmsford Formation which are subparallel to the long axis of the Sudbury Basin. These authors suggested that the wackes of the Chelmsford Formation were deposited in an elongate trough and not in a circular basin. However, they did not take into account the rotation of paleocurrent indicators due to strain. Rotation due to folding, however, was considered. Most importantly, these authors and Brocoum and Dalziel (1974) did not take into account that destraining of the sedimentary core of the Sudbury Structure is not equivalent to destraining the entire Sudbury Structure (Shanks 1991). Large parts of the Structure do not exhibit any evidence of

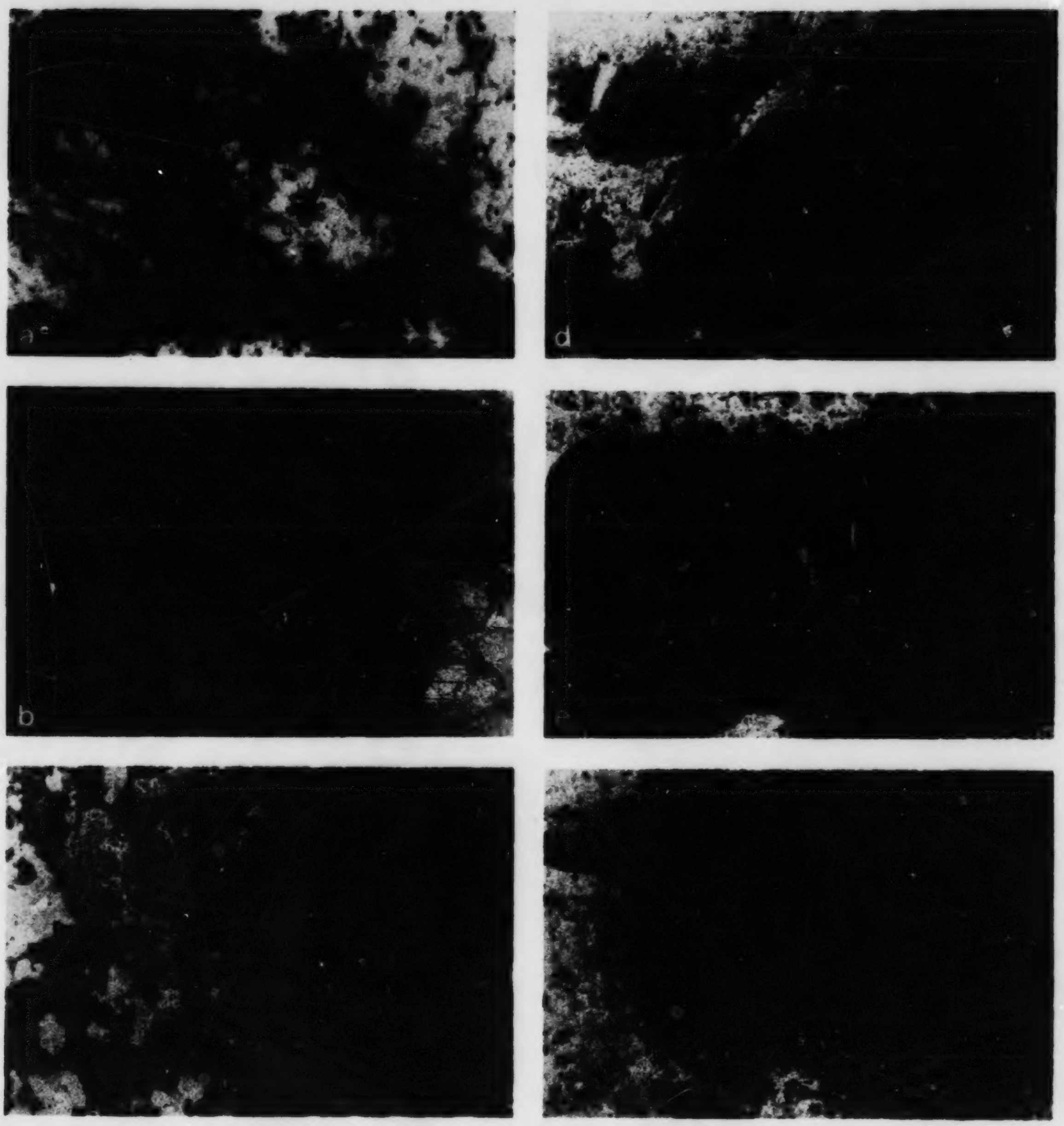


Figure 15.17. Several ore samples from Sudbury mines. a) Massive sulphide stringers (pyrrhotite and pentlandite, minor chalcopyrite) containing Levack Gneiss fragment. Levack Mine. Length of photographed rock slab 7.5 cm. b) Massive sulphides (coarse-grained, twinned chalcopyrite and pentlandite along some twin planes, rich in total precious metals). Little Stobie Mine. Length of photographed slab is 7.0 cm. c) Sulphide (pyrrhotite with small patches of pentlandite) blebs in quartz diorite matrix. Frood Mine. Length of photographed slab is 7.0 cm. d) Fine-grained fragments of mafic metavolcanic rock rimmed and intruded by massive sulphide (heterogeneous mixture of pyrrhotite, chalcopyrite, and pentlandite). Stobie Mine. Length of photographed slab is 7.0 cm. e)Deformed sulphide (mainly chalcopyrite with some arsenide phases near the top of the photograph, namely gersdorffite and probable intergrowth of maucherite and niccolite. This assemblage is usually high in PGEs) and clasts. Vermilion Mine. Length of photographed slab is 11.8 cm. f) Massive sulphide (heterogeneous assemblage of pyrrhotite and chalcopyrite, minor pentlandite) with aligned, deformed, fine-grained mafic inclusions. Garson Mine. Length of photographed slab is 7.5 cm. (Photography and sulphide determinations *courtesy of* G.G.Morrison, Inco Limited, Copper Cliff).

strain. Muir (1984) compared the present shape of the Sudbury Structure with a hypothetical, circular, predeformation structure and listed several reasons for non-initial circularity. In particular, he cited a lack of deformation features north of the Sudbury Igneous Complex (up to at least 30 km away) which would corroborate the required deformation for transforming an initial circular structure into the present elliptical one.

GEOPHYSICAL CHARACTERISTICS OF THE SUDBURY STRUCTURE

Gravity Anomalies

A combined interpretation of data from gravity and magnetic surveys of the Sudbury Structure was recently carried out by Gupta et al. (1984). An examination of the anomalous Bouguer gravity map with the regional field removed (Figure 15.20), and the total field magnetic map (Figure 15.21), reveals that both gravity and magnetic anomalies in the area are much broader than the Sudbury Structure itself. The Sudbury Structure occurs within a large, oval-shaped gravity high of about +30 mgal that is part of a 350 km long, chain-like zone of positive gravity anomalies, extending from Elliot Lake to Englehart. Another notable feature is that the Sudbury Structure does not lie along the axis of the gravity high, but is offset to the north by about 8 to 10 km.

Intense, but more local, positive gravity anomalies occur along a line that parallels the southern boundary of the Sudbury Structure (*see* Figure 15.20). These are the Agnew Lake gravity high which coincides with outcropping anorthositic gabbro, and the Victoria Mine and Lively gravity highs under the Elsie Mountain and Stobie Formations. Gravity lows coincide with the Killarney Granite Pluton, the Creighton Granite, the Venetian Lake granite pluton and Lake Wanapitei. The small Benny Greenstone Belt is outlined by a weaker but definite gravity high. Another prominent gravity high occurs over the anorthosite gabbro complex near River Valley northeast of Sudbury.

Density Anomalies

The ability to measure gravity anomalies depends upon the contrast between the densities of different rock units and the background density. Rock densities for individual rock units of the Sudbury Structure and footwall rocks were summarized by Gupta et al. 1984. To compute the gravity effect of the Sudbury Structure, a model value for the background density of the rocks that lie beneath, and surround, the structure must be established (Gupta et al. 1984). A weak positive anomaly is associated with the Levack Gneiss (average density 2.77 x 10^3 kg/m^3) and weak gravity lows are associated with sandstones and wackes of the Huronian Supergroup

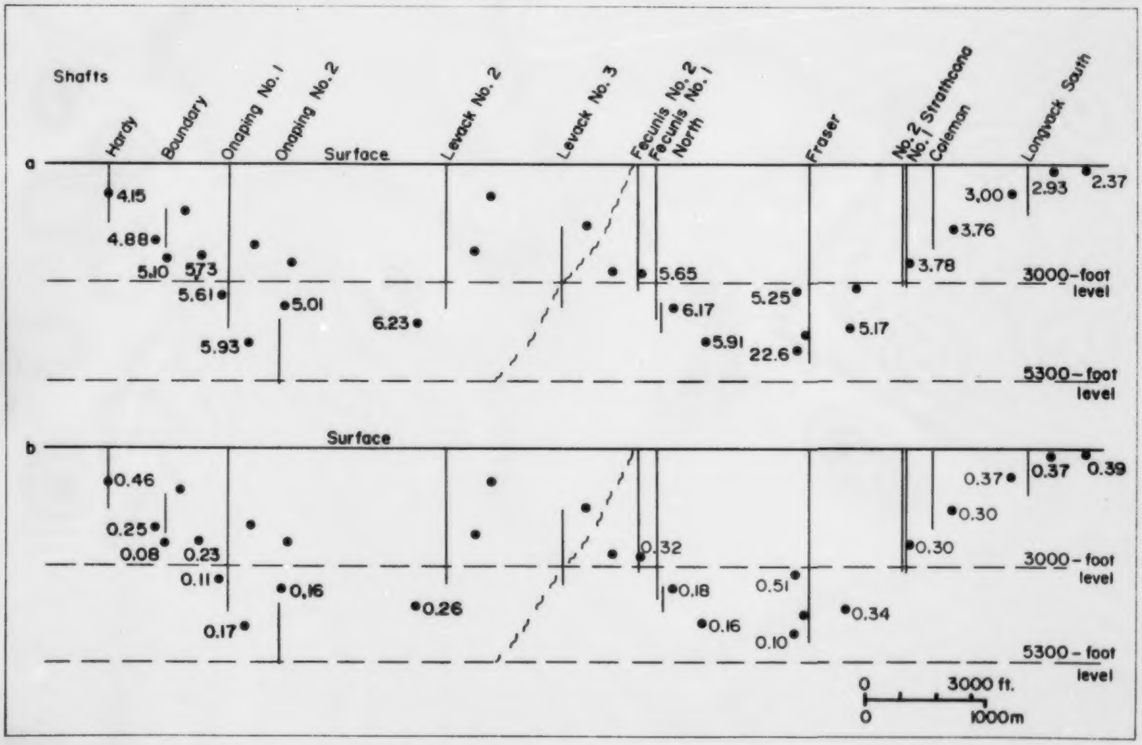

Shafts
Hardy
Boundary
Onaping No. 1
Onaping No. 2
Levack No. 2
Levack No. 3
Fecunis No. 2
Fecunis No. 1
North
Fraser
No. 2 Strathcona
No. 1
Coleman
Longvack South
a
Surface
4.15
4.88
5.10
5.73
5.61
5.93
5.01
6.23
5.65
6.17
5.91
5.25
22.6
5.17
3.78
3.76
3.00
2.93
2.37
3000-foot level
5300-foot level
b
Surface
0.46
0.25
0.08
0.23
0.11
0.17
0.16
0.26
0.32
0.18
0.16
0.51
0.10
0.34
0.30
0.30
0.37
0.37
0.39
3000-foot level
5300-foot level
0
3000 ft.
0
1000m

Figure 15.18. Diagrammatic, vertical, longitudinal section through the mines of the Onaping Levack area. Dots represent approximate central locations of individual ore zones. a) Percent nickel in sulphides. b) Ratios of Cu/Cu+Ni. (*from* Coats and Snajdr 1984).

(average density 2.66 x 10^3 kg/m^3). The background density should therefore lie between 2.66 x 10^3 and 2.77 x 10^3 kg/m^3. The region that lies immediately north of the Sudbury Igneous Complex is gravitationally flat, indicating that the gneissic rocks (average density 2.73 x 10^3 kg/m^3) in this area have a density that is close to the background value. Thus, the most rational choice for the background density is 2.73 x 10^3 kg/m^3.

Modelling

The Bouguer gravity field of the Sudbury Structure area was divided into regional and residual anomalous Bouguer gravity components (*see* Figure 15.20). The residual anomalous Bouguer gravity field was further separated into "major" (regional) and "minor" (residual) components by Gupta et al. (1984). The regional component of the residual anomalous Bouguer gravity was termed as the "Sudbury Gravity Anomaly." This anomaly has been expressed as being due to either:

1. two overlapping positive single peaked anomalies, called the northern and southern anomalies, caused by a small and a large body, respectively (Model A)
2. a combination of a single positive anomaly with a small negative anomaly superimposed on it (Model B)

MODEL A

A least square inversion, using finite strike length models of polygonal cross sections, shows a large, dense mass underlying, and extending well to the south of, the South

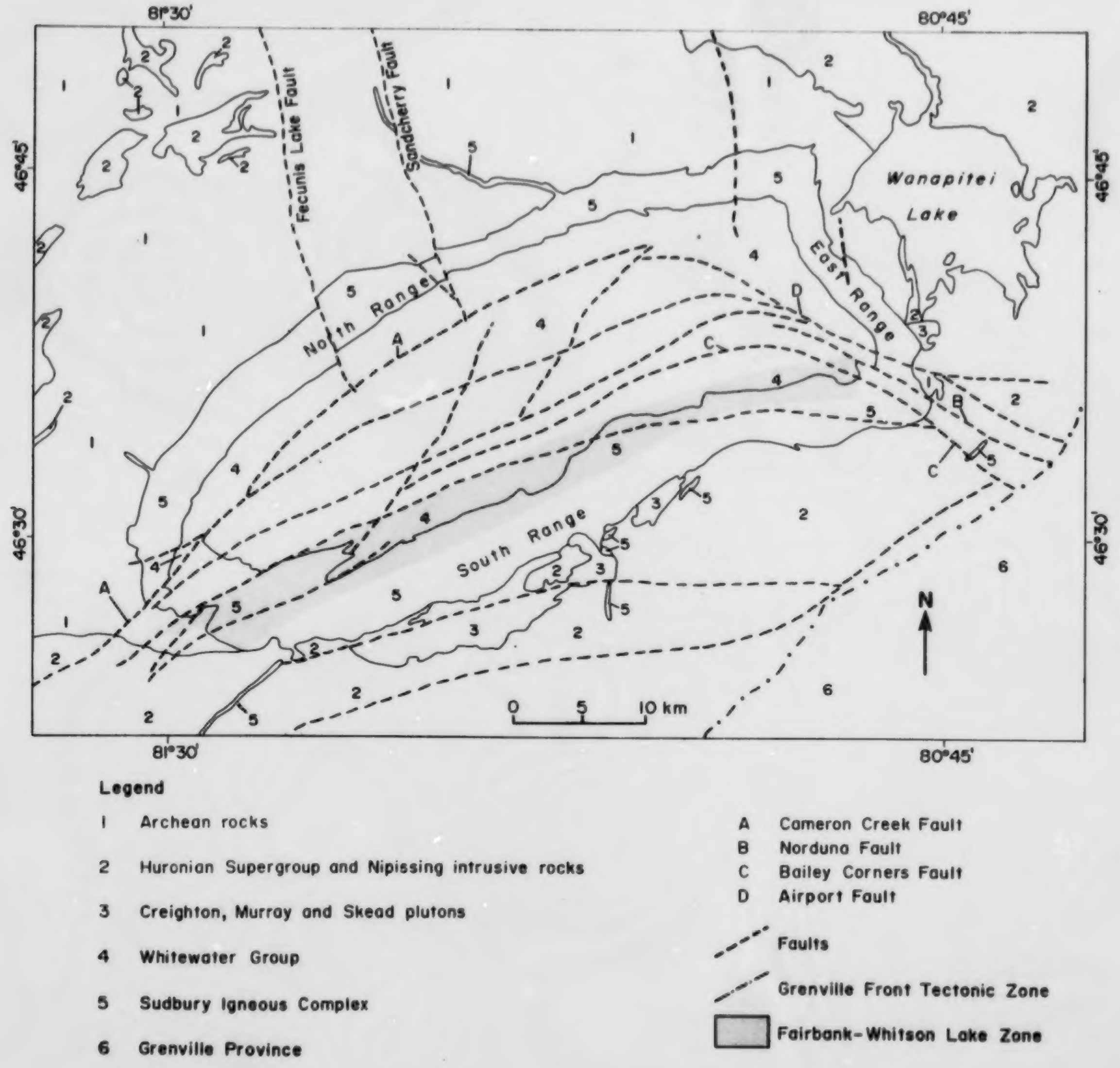

81°30'
80°45'
46°45'
46°30'
Fecunis Lake Fault
Sandcherry Fault
North Range
East Range
South Range
Wanapitei Lake
A
B
C
D
N
0
5
10 km
Legend
1 Archean rocks
2 Huronian Supergroup and Nipissing intrusive rocks
3 Creighton, Murray and Skead plutons
4 Whitewater Group
5 Sudbury Igneous Complex
6 Grenville Province
A Cameron Creek Fault
B Norduna Fault
C Bailey Corners Fault
D Airport Fault
Faults
Grenville Front Tectonic Zone
Fairbank-Whitson Lake Zone

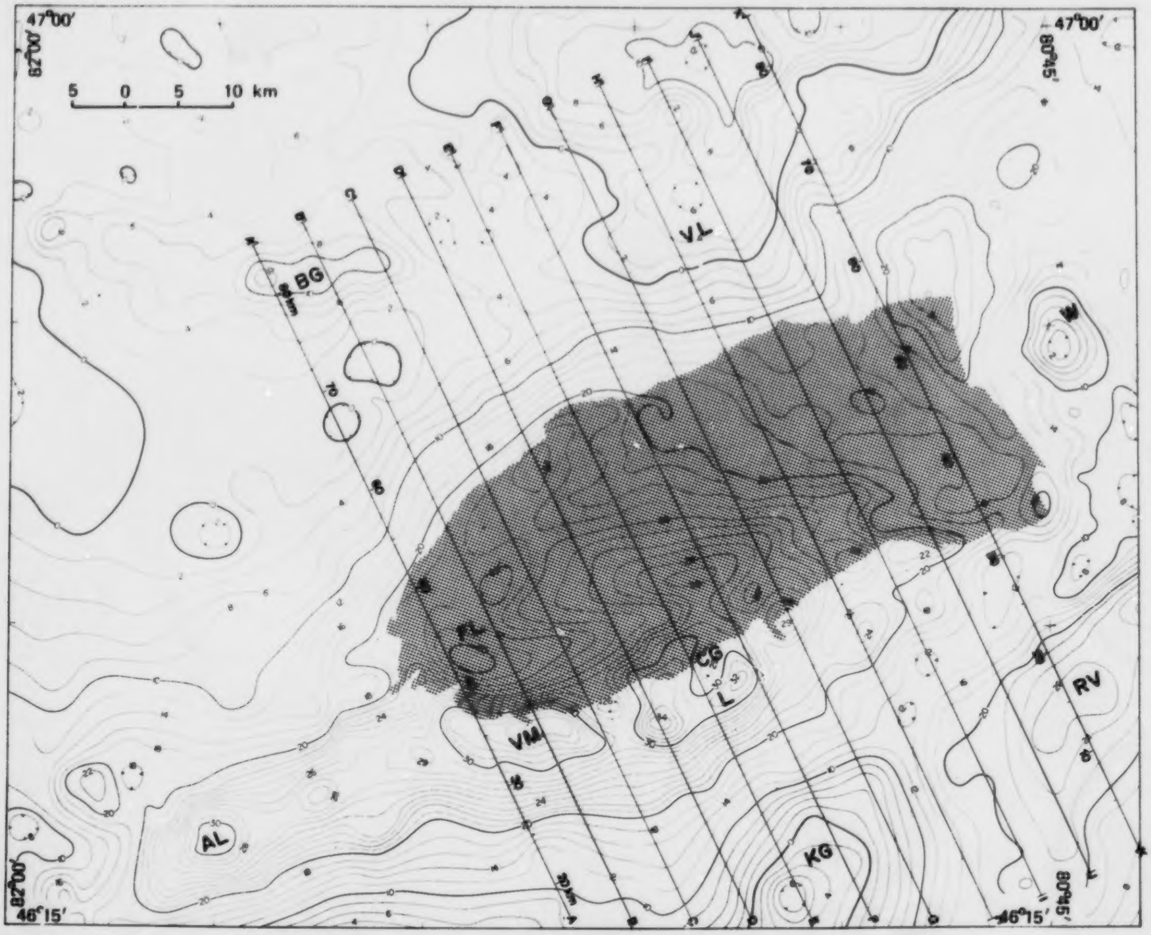
Figure 15.19. Structural geology of the Sudbury Structure (in part *after* Shanks 1991 and Cowan and Schwerdtner 1990).

Range (*see* profile EE′, Figure 15.22). It is sill-like in form with depth to its top between 6 and 9 km. The smaller mass underlying the North Range is more dike-like, having a horizontal thickness of about 4 km, and is buried to a depth of about 3 km. It is postulated that the rocks of this dike-like mass are mafic to ultramafic in composition, with a mineralogy resembling anorthositic gabbro, olivine gabbro or dunite. These are probably the only rock types with a density as high as that suggested by modelling (3.08×10^3 kg/m^3 for the large body; 3.02×10^3 kg/m^3 for the small body). The mineralogy of the large and small bodies must be different, since the shallow small body is strongly magnetic compared to the deeper large body which is only weakly magnetic.

MODEL B

A least square inversion of the positive gravity anomaly, again, suggests a broad sill-like mass as the cause of the anomaly (*see* profile EE′, Figure 15.23). Depth to its top is between 4 and 6 km for a density contrast of $+0.29 \times 10^3$ kg/m^3. The rock type thus appears to be the same as obtained by using the previous model, namely, anorthosite or olivine gabbro or dunite. The depth to the top of the body is shallower in this model because of the assumption that the entire positive anomaly is caused by a single body.

The inversion of the negative anomaly was carried out on the assumption that it is caused by rocks within the



AL= Agnew Lake Anorthosite Gabbro FL= Fairbank Lake granophyre VL= Venetian Lake granite pluton

BG= Benny greenstone belt KG= Killarney granite pluton RV= River Valley anorthosite complex

CG= Creighton granite L= Lively gravity high VM= Victoria Mine gravity high W= Lake Wanapitei

Figure 15.20. Anomalous Bouguer gravity field. Location of profiles is also shown (*from* Gupta et al. 1984).

Sudbury Structure that have lower densities than the background value. A two-layer model of the Sudbury Basin rocks was therefore proposed. The outer layer of the negative mass (i.e., lighter than the background density) consists of granophyre and Onaping Formation rocks with a combined average density of 2.71 x 10^3 kg/m^3, against a background density of 2.73 x 10^3 kg/m^3 (density contrast -0.02 x 10^3 kg/m^3). The inner layer consists of the Onwatin and Chelmsford formations with a combined average density of 2.68 x 10^3 kg/m^3 (density contrast -0.05 x 10^3 kg/m^3). During computer modelling, the density contrast and the surface limits of the 2 layers were kept fixed at their geologically mapped positions. A depth of about 4.5 km was obtained to the base of the basin-like granophyre body .

Aeromagnetic Anomalies

The Sudbury Structure is defined by a distinct elliptical positive magnetic anomaly interpreted to be caused by 3 major magnetic sources lying at different depths. These sources are:

1. a semicontinuous elliptical magnetic anomaly, probably caused by surface to near-surface oxide-rich gabbros of the Sudbury Igneous Complex. A susceptibility of 0.19 to 0.25 SI units has been used in computer modelling (*see* Figures 15.22 and 15.23).
2. a linear narrow magnetic anomaly occurring close to the contact between the Onaping and Onwatin formations on the south side of the Sudbury Basin. This anomaly is probably related to the gabbroic rocks intruding the rocks of the Whitewater Group. A susceptibility of between 0.03 to 0.08 SI units has been used in computer modelling (*see* Figures 15.22 and 15.23).
3. a broad regional anomaly which reaches a maximum amplitude of 1000 nT near the North Range (Figure 15.24) is perhaps due to rocks which are buried and occur at depth. The anomaly is caused by magnetic sources at two different depths. The first is a large tabular magnetic mass (magnetic susceptibility 0.25 SI units) dipping at about 30°S buried at a depth of about 4 km (*see* profile EE', Figures 15.22 and 15.23). The second, steeply dipping, smaller plate-like body (apparent

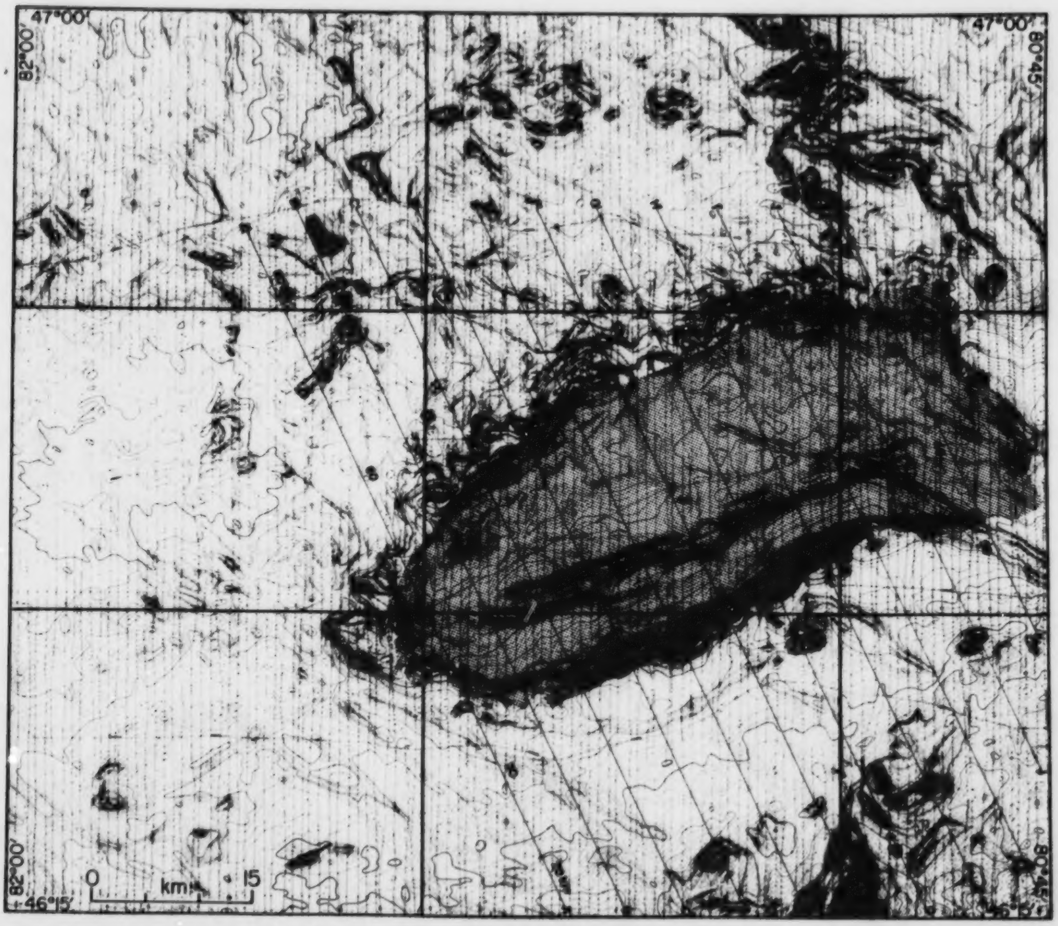


Figure 15.21. Total-field aeromagnetic contours. Profile locations are also shown. (*from* Gupta et al. 1984).

susceptibility 0.8/t SI units where "t" is the thickness in kilometres) is buried to a depth of about 9 km.

Discussion

The combination of high density and high apparent magnetic susceptibility, yielded from inversion process, suggests that a large volume of mafic to ultramafic rocks, which is not exposed at the surface, lies beneath the Sudbury Structure. However, these rocks may be related to gabbroic and ultramafic rocks that appear locally as inclusions in the Sublayer. The buried rocks are at least partly serpentinized, since this is the only process that can lead to such strong magnetization in dense ultramafic rocks. The mafic and ultramafic inclusions in the Sublayer, however, are not serpentinized. The hidden sill-like body is subhorizontal in attitude and extends beyond the boundaries of the Sudbury Igneous Complex in all directions. Its southern part is buried to a depth of between 5 and 8 km, while the northern part of this body (or a separate, smaller body) lies at a depth of between 2 and 5 km.

Data from magnetic surveys support a gravitationally neutral Sudbury Basin underlain by 2 disconnected bodies, one large and one small, having densities similar to gabbro or gabbro-anorthosite (3×10^3 to 3.3×10^3 kg/m^3), one of them serpentinized. However, geological mapping supports a gravitationally negative Sudbury Basin underlain by a single sill-like mass, of which only the northern, raised portion is serpentinized.

THE ORIGIN OF THE SUDBURY STRUCTURE

Until the early 1960s the Sudbury Structure was considered to have an endogenic, volcanic or volcanic-tectonic origin (Burrows and Rickaby 1930; Speers 1957; Stevenson 1961b; Thomson 1957; Williams 1957). In a landmark paper, Dietz (1964) suggested that the Sudbury Structure

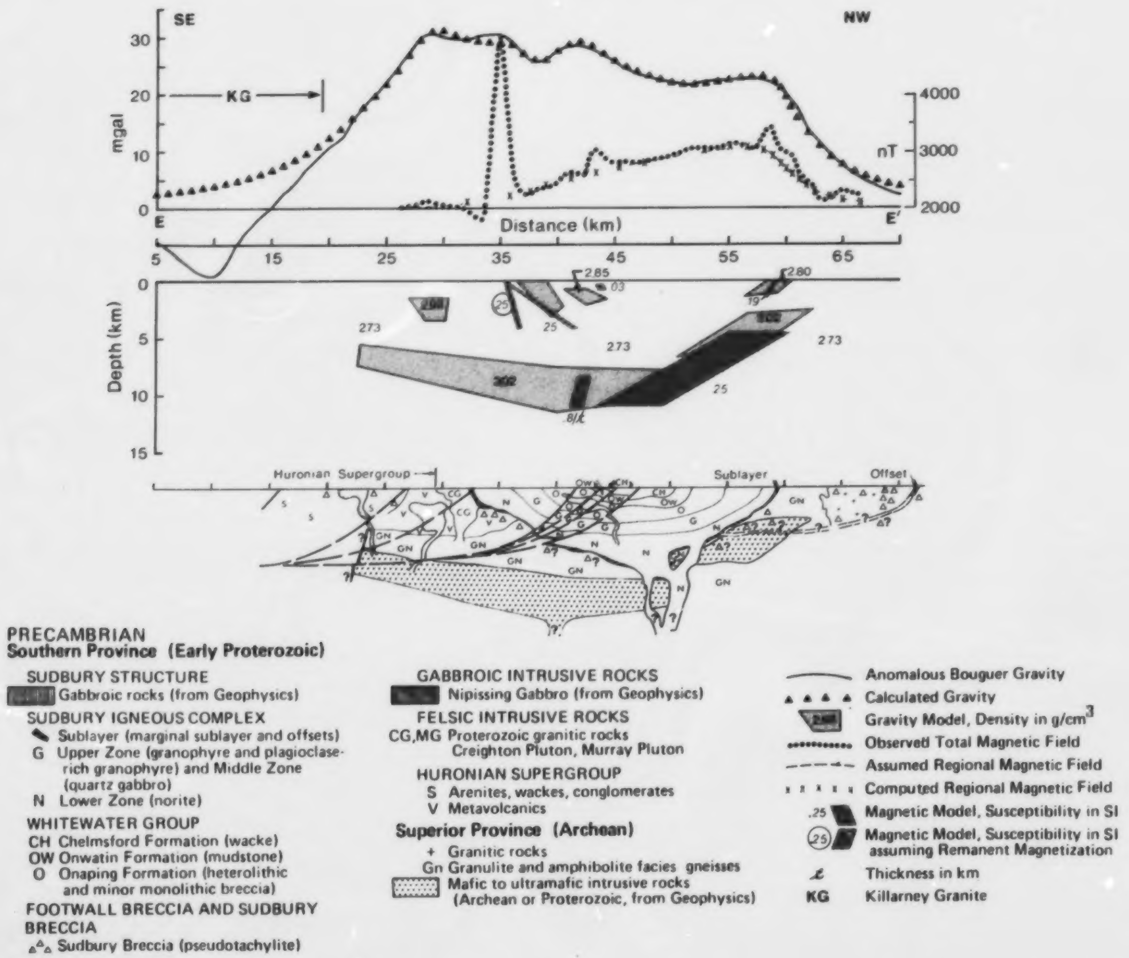

SE
NW
mgal
30
20
10
0
KG
nT
4000
3000
2000
E
E'
Distance (km)
5
15
25
35
45
55
65
Depth (km)
0
5
10
15
2.85
2.80
.03
.19
25
2.73
2.73
2.73
25
3.02
3.02
.8/ℓ
25
Huronian Supergroup
Sublayer
Offset
S
V
CG
G
N
GN
OW
CH
O
Δ?
PRECAMBRIAN
Southern Province (Early Proterozoic)
SUDBURY STRUCTURE
Gabbroic rocks (from Geophysics)
SUDBURY IGNEOUS COMPLEX
Sublayer (marginal sublayer and offsets)
G Upper Zone (granophyre and plagioclase-rich granophyre) and Middle Zone (quartz gabbro)
N Lower Zone (norite)
WHITEWATER GROUP
CH Chelmsford Formation (wacke)
OW Onwatin Formation (mudstone)
O Onaping Formation (heterolithic and minor monolithic breccia)
FOOTWALL BRECCIA AND SUDBURY BRECCIA
ΔΔ Sudbury Breccia (pseudotachylite)
GABBROIC INTRUSIVE ROCKS
Nipissing Gabbro (from Geophysics)
FELSIC INTRUSIVE ROCKS
CG,MG Proterozoic granitic rocks Creighton Pluton, Murray Pluton
HURONIAN SUPERGROUP
S Arenites, wackes, conglomerates
V Metavolcanics
Superior Province (Archean)
+ Granitic rocks
Gn Granulite and amphibolite facies gneisses
Mafic to ultramafic intrusive rocks (Archean or Proterozoic, from Geophysics)
Anomalous Bouguer Gravity
Calculated Gravity
Gravity Model, Density in g/cm3
Observed Total Magnetic Field
Assumed Regional Magnetic Field
Computed Regional Magnetic Field
.25 Magnetic Model, Susceptibility in SI
(25) Magnetic Model, Susceptibility in SI assuming Remanent Magnetization
ℓ Thickness in km
KG Killarney Granite

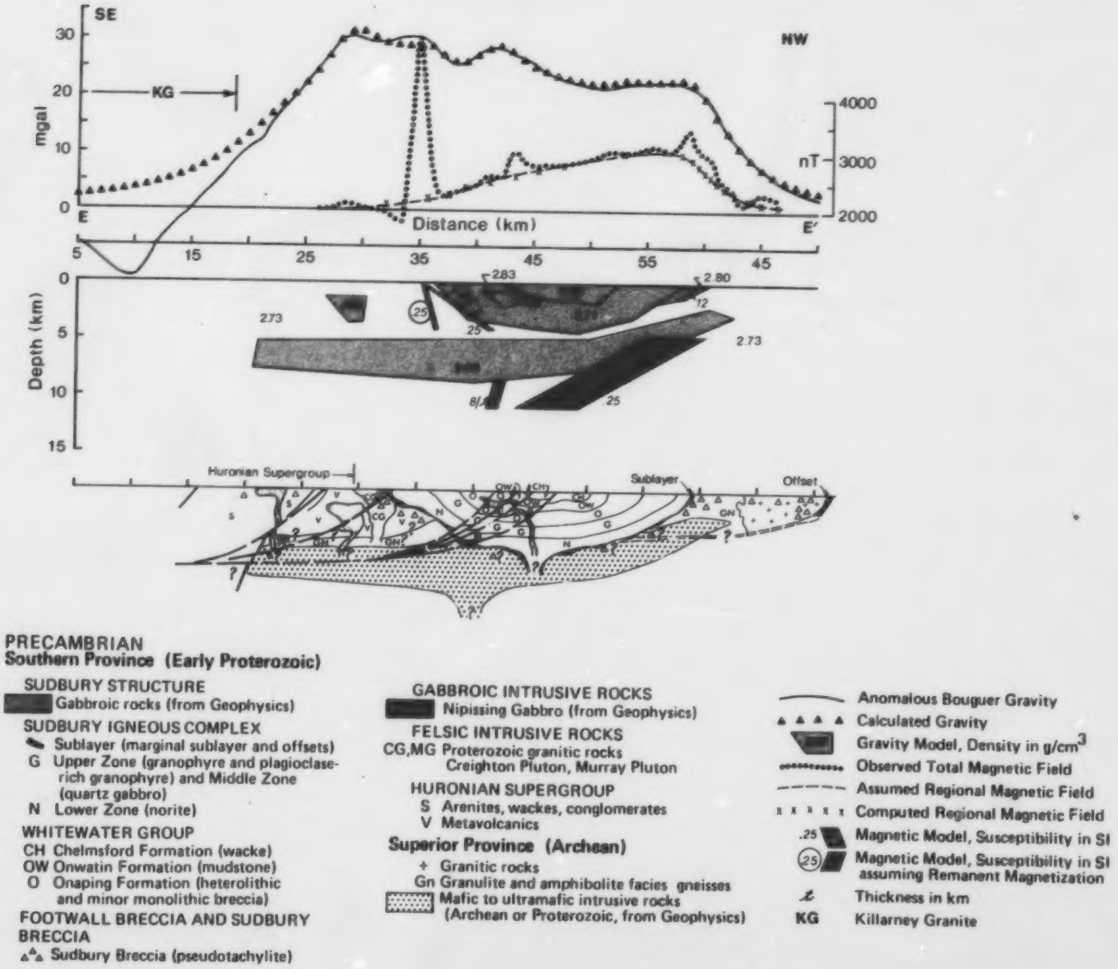
Figure 15.22. Profile EE′ showing the composite gravity interpretation of Sudbury gravity Model A and residual anomalies. Computed model for profile EE′ is also shown. (*from* Gupta et al. 1984).

represented a meteorite impact structure and that the heterolithic breccias of the Onaping Formation were formed by explosive degassing over an ascending magma whose intrusion had been triggered by the impact. Impact-induced volcanism was also suggested by Thomson (1969) and Muir (1982, 1983). Many papers have been published favouring an impact origin for the Sudbury Structure including Dence (1972), Dietz (1964), Dietz and Butler (1964), Dressler (1984c), Dressler et al. (1987), French (1968, 1970, 1972), Guy-Bray et al. (1966), Morrison (1984), Peredery (1972a, 1972b), Peredery and Morrison (1984), and Stöffler et al. (1989). Advocates of an endogenic origin likewise are numerous, Cantin and Walker (1972), Card and Hutchinson (1972), Fleet (1979, 1980), Muir (1984), Stevenson (1972), and Stevenson and Stevenson (1980).

In Tables 15.3 and 15.4, the authors list evidence in support of both an impact origin (Table 15.3)and an endogenic origin (Table 15.4). These 2 lists are not complete; they do not contain all aspects ever published in favour of, or against, the 2 concepts. The authors are aware that there are counter-arguments to the arguments listed in both tables. Detailed discussions on the origin of the Sudbury Structure were published in Pye et al. (1984), by Morrison (1984), Muir (1984), Naldrett (1984c), and Peredery and Morrison (1984). Dressler et al. (1987) also discussed the origin of the Sudbury Structure.

It is beyond the scope and purpose of this review to discuss the various arguments listed in Tables 15.3 and 15.4. The central issue regarding the origin of the Sudbury Structure, however, is worth reiterating and is as follows.

The presence of shock metamorphic features in the Onaping Formation and in the footwall rocks of the Sudbury Structure, and the extreme brecciation of country rocks around the Sudbury Igneous Complex are evidence of a catastrophic event. A meteorite impact can account for the observed shock metamorphic features and the brecciation.


SE
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mgal
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KG
E
E′
Distance (km)
4000
nT
3000
2000
5
15
25
35
45
55
45
Depth (km)
0
5
10
15
283
2.80
12
2.73
25
2.73
25
8/4
25
Huronian Supergroup
Sublayer
Offset
S
V
CG
G
N
OW
CH
O
GN
PRECAMBRIAN
Southern Province (Early Proterozoic)
SUDBURY STRUCTURE
Gabbroic rocks (from Geophysics)
SUDBURY IGNEOUS COMPLEX
Sublayer (marginal sublayer and offsets)
G Upper Zone (granophyre and plagioclase-rich granophyre) and Middle Zone (quartz gabbro)
N Lower Zone (norite)
WHITEWATER GROUP
CH Chelmsford Formation (wacke)
OW Onwatin Formation (mudstone)
O Onaping Formation (heterolithic and minor monolithic breccia)
FOOTWALL BRECCIA AND SUDBURY BRECCIA
Sudbury Breccia (pseudotachylite)
GABBROIC INTRUSIVE ROCKS
Nipissing Gabbro (from Geophysics)
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CG,MG Proterozoic granitic rocks Creighton Pluton, Murray Pluton
HURONIAN SUPERGROUP
S Arenites, wackes, conglomerates
V Metavolcanics
Superior Province (Archean)
+ Granitic rocks
Gn Granulite and amphibolite facies gneisses
Mafic to ultramafic intrusive rocks (Archean or Proterozoic, from Geophysics)
Anomalous Bouguer Gravity
Calculated Gravity
Gravity Model, Density in g/cm3
Observed Total Magnetic Field
Assumed Regional Magnetic Field
Computed Regional Magnetic Field
.25 Magnetic Model, Susceptibility in SI
25 Magnetic Model, Susceptibility in SI assuming Remanent Magnetization
Thickness in km
KG Killarney Granite

Figure 15.23. Profile EE′ showing the gravity interpretation consisting of Sudbury gravity anomaly Model B and residual anomalies. Near-surface models computed from magnetic anomalies for profile EE′ are also shown. (*from* Gupta et al. 1984).

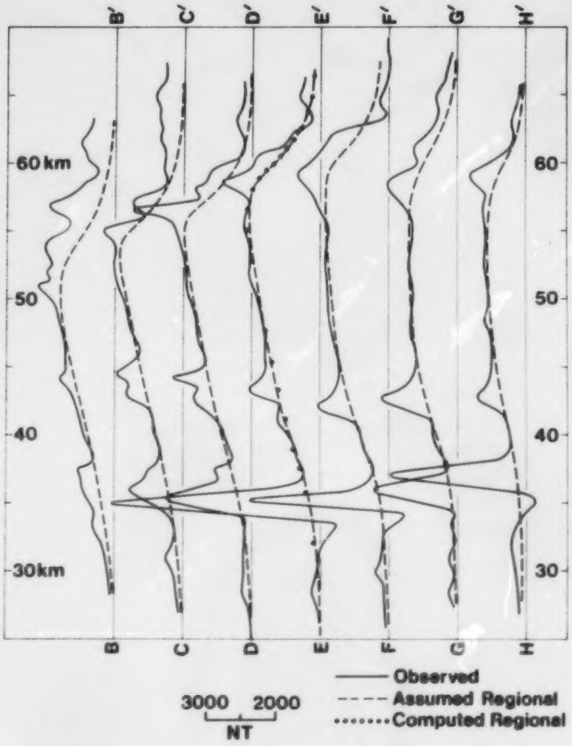

Figure 15.24. Separation of total magnetic intensity into regional and residual components.

These features have not been found in any undisputed endogenic feature in the Canadian Shield or elsewhere on this planet. Field observations such as those listed under (1) in Table 15.4, however, are evidence of multiple processes. Many of these are not in agreement with impact processes as presently understood.

Only unbiased observation, co-operation and new research initiatives will advance our understanding of the Sudbury Structure. The authors welcome the deep crustal seismic investigations of the Canadian LITHOPROBE program in Sudbury as new initiatives. Deep continental drilling and computer modelling of endogenic and exogenic explosions, and excavation and deposition processes are also needed to bring science closer to an understanding of the Sudbury Structure. Comparative planetology may also play a major role in future research.

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Chapter 16

Proterozoic Geology of the Lake Superior Area

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Abstract

Proterozoic rocks of Ontario in the Lake Superior area form cover sequences on and are intrusive into Archean rocks of the Superior Province. These rocks provide evidence of 3 distinct events in the Proterozoic; however, the dominant tectonic event in the region was the development of the Midcontinent Rift along the present axis of Lake Superior.

In the Thunder Bay area, the Paleoproterozoic Animikie Group forms a south-facing homoclinal sequence consisting of iron formation, chert and argillite of the Gunflint Formation and overlying shales and greywacke of the Rove Formation. The relatively undeformed, clastic sedimentary rocks of the Animikie Group are interpreted to have been largely derived from 1.86 billion-year-old rocks in the Penokean Orogen to the south of Lake Superior.

South of Lake Nipigon, the Mesoproterozoic Sibley Group sedimentary rocks form a thin, undeformed, epicontinental, clastic, sedimentary sequence, which unconformably overlies Archean and Animikie Group rocks. The Sibley Group, which is now largely preserved in a graben extending north from Lake Superior to Lake Nipigon, largely consists of quartz arenite, argillaceous dolomite and mudstones which are interpreted to have been deposited in a playa lake environment. In the Lake Nipigon area, Mesoproterozoic anorogenic alkali granite, dated at 1.54 Ga, is overlain by minor volcanic rocks and quartz arenite; the latter is probably correlative with the basal part of the Sibley Group.

The formation of the Midcontinent Rift at 1.11 to 1.09 Ga resulted in the deposition of the Keweenawan Supergroup. Along the axis of Lake Superior, the rocks of the Keweenawan Supergroup were deposited in asymmetric grabens bounded by normal faults which were subsequently reactivated as reverse faults. The lower and middle parts of the supergroup are dominated by volcanic rocks and consist of subaerial tholeiitic flows, felsic volcanic rocks and minor fluvial sedimentary rocks. Related intrusive rocks include layered gabbro plutons, diabase sills and dike swarms, mafic to felsic ring complexes and zoned plutons. The upper part of the Keweenawan Supergroup consists dominantly of quartz arenite which unconformably overlies the volcanic part of the sequence. The entire Keweenawan Supergroup reaches a maximum thickness of approximately 30 km beneath Lake Superior. Carbonatite and alkalic complexes of Keweenawan age are associated with zones of faulting in the Superior Province north of Lake Superior.

Mineral deposits in the Proterozoic rocks of the Lake Superior area in Ontario include: copper, nickel and platinum group element (PGE) mineralization in layered gabbroic plutons; copper deposits in veins and amygdules in Keweenawan basalts; silver mineralization in calcite-barite-quartz veins in diabase and Animikie Group sedimentary rocks; uranium mineralization in fractures in Archean rocks near the Proterozoic unconformities; and amethyst veins present in Archean rocks that have been fractured during the Proterozoic. There is currently only production from amethyst deposits; however, there has been past production from silver deposits, such as the Silver Islet Mine, and copper deposits in the Mamainse Point area.

INTRODUCTION

This chapter gives an overview of the Proterozoic geology of Ontario in the area north of Lake Superior. This part of the province provides a geological record of a long history of Proterozoic events which culminated in the development of the Midcontinent Rift, one of the largest and deepest continental rift structures preserved on the earth. No traveller through this area will forget the bold and picturesque topography along the North Shore of Lake Superior, and much of this is a result of tectonic and magmatic events related to this rift.

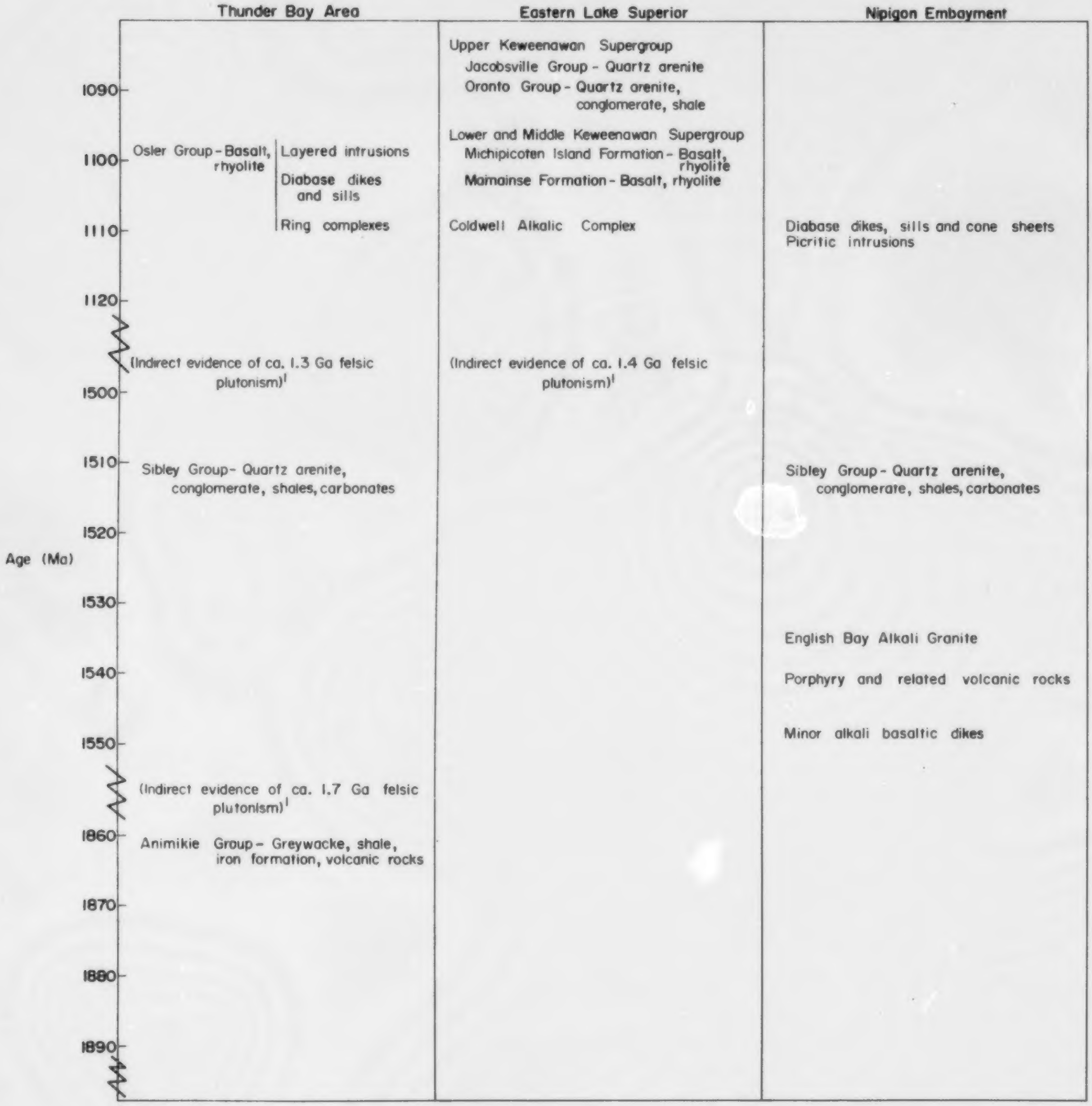
The Proterozoic rocks of the Lake Superior area within Ontario form part of the Southern Province of the Canadian Shield (Stockwell et al. 1972). Evidence of at least 3 distinct events in the Paleoproterozoic and Mesoproterozoic (Figures 16.1 and 16.2) is preserved. The earliest of these events is the deposition of Paleoproterozoic Animikie Group sediments prior to and during the Penokean Orogeny at ~1.86 Ga (Sims et al. 1989). During a period of relative tectonic stability in the Lake Superior region, intrusion of Mesoproterozoic anorogenic granites at 1.54 Ga and subsequent deposition of the Sibley Group occurred north of Lake Superior. Finally, the dominant tectonic event in this region was Mesoproterozoic rifting at 1.11 to 1.09 Ga during which time the Keweenawan Supergroup was deposited in the Midcontinent Rift (Cannon et al. 1989).

In Ontario, north of Lake Superior, Proterozoic supracrustal rocks form 1 to 2 km thick remnants of intracratonic cover sequences on Archean rocks of the Superior Province. Archean basement is also inferred to be present beneath the Midcontinent Rift in Lake Superior, despite accumulations of up to 30 km of rift-filling volcanic rocks and sedimentary rocks that are indicated by seismic reflection profiles

(Cannon et al. 1989). Proterozoic rocks that represent juvenile crust were accreted onto the southern margin of the Superior Province during the ~1.86 billion-year-old Penokean Orogeny and in the ~1.1 billion-year-old Grenville Orogeny. The juvenile Proterozoic crust of Penokean age, south of Lake Superior, has been recently described by Sims et al. (1989) and Barovich et al. (1989); the Grenville Province is reviewed by Easton (this volume).

BRIEF HISTORICAL BACKGROUND OF GEOLOGICAL STUDIES IN THE LAKE SUPERIOR AREA

The geology of the region was initially discussed in a review of the Precambrian geology of Canada by Sir William



Age (Ma)	Thunder Bay Area	Eastern Lake Superior	Nipigon Embayment
		Upper Keweenawan Supergroup Jacobsville Group - Quartz arenite Oronto Group - Quartz arenite, conglomerate, shale	
1090			
1100	Osler Group - Basalt, rhyolite Layered intrusions Diabase dikes and sills	Lower and Middle Keweenawan Supergroup Michipicoten Island Formation - Basalt, rhyolite Mamainse Formation - Basalt, rhyolite	
1110	Ring complexes	Coldwell Alkalic Complex	Diabase dikes, sills and cone sheets Picritic intrusions
1120			
	(Indirect evidence of ca. 1.3 Ga felsic plutonism)[1]	(Indirect evidence of ca. 1.4 Ga felsic plutonism)[1]	
1500			
1510	Sibley Group - Quartz arenite, conglomerate, shales, carbonates		Sibley Group - Quartz arenite, conglomerate, shales, carbonates
1520			
1530			
			English Bay Alkali Granite
1540			Porphyry and related volcanic rocks
			Minor alkali basaltic dikes
1550			
	(Indirect evidence of ca. 1.7 Ga felsic plutonism)[1]		
1860	Animikie Group - Greywacke, shale, iron formation, volcanic rocks		
1870			
1880			
1890			

Notes

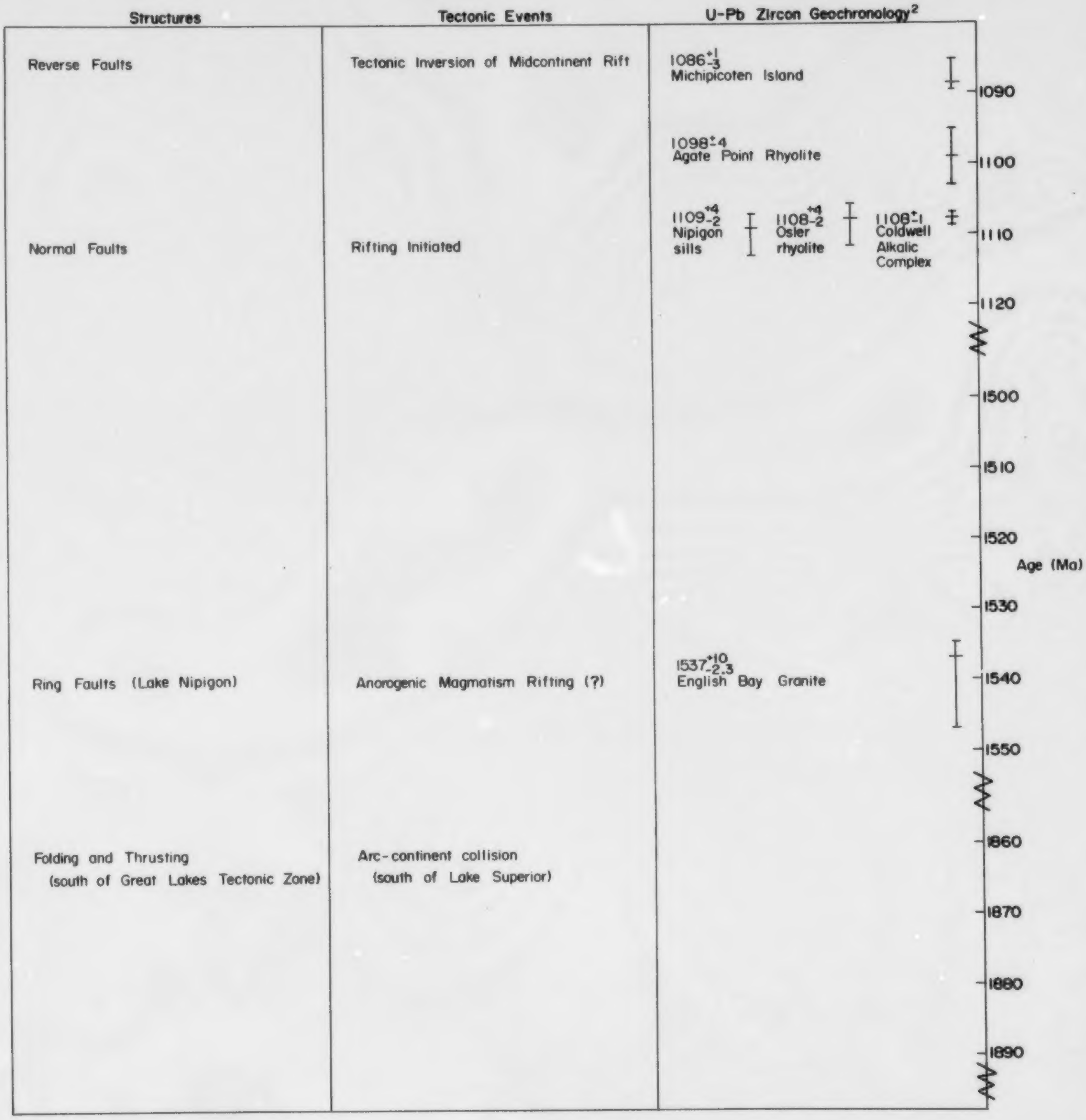
1. Based on dating of clasts (Davis and Sutcliffe 1985) and inherited zircons (Palmer and Davis 1987).

Figure 16.1. Time chart for geological events related to the Proterozoic evolution of the Lake Superior area in Ontario. Specific dike swarms and alkalic complexes (except the Port Coldwell alkalic complex) are not included in this figure.

Logan, who in 1863, recognized an "Upper Copper-bearing series" that was believed to be stratigraphically above the Huronian rocks north of Lake Huron (Logan 1863).

One of the first surveys in the northern Lake Superior area was initiated in 1860, when the Crown Lands Department of Ontario sent J.W. Herrick to run a colonization line around Lake Superior and explore the country. This work took 3 years, beginning near Sault Ste. Marie and ending on the Kaministiquia River near Thunder Bay. Herrick's work forms the basis of most subsequent surveys. Production at the Silver Islet Mine, near Thunder Bay, began in 1869. The Silver Islet Mine became one of Ontario's most significant mineral producers.

Scientific surveys followed the early mineral discoveries and geographic surveys around 1869, when Robert Bell of the Geological Survey of Canada was dispatched to make



2. See text for references.

Some diabase dike swarms have not been included for simplification of the table.

Figure 16.1. Time chart for geological events related to the Proterozoic evolution of the Lake Superior area in Ontario. Specific dike swarms and alkalic complexes (except the Port Coldwell alkalic complex) are not included in this figure.

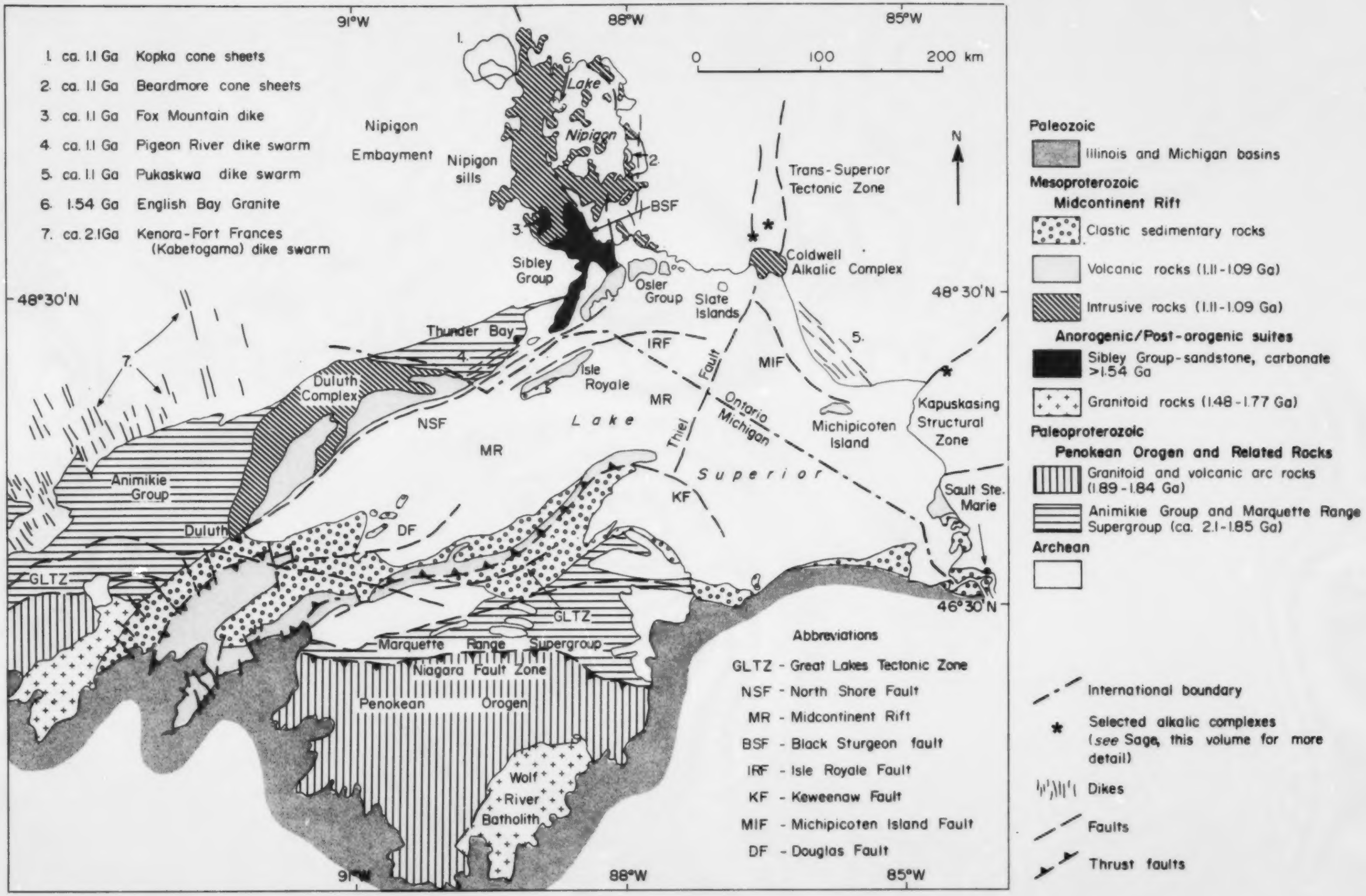

1. ca. 1.1 Ga Kopka cone sheets
2. ca. 1.1 Ga Beardmore cone sheets
3. ca. 1.1 Ga Fox Mountain dike
4. ca. 1.1 Ga Pigeon River dike swarm
5. ca. 1.1 Ga Pukaskwa dike swarm
6. 1.54 Ga English Bay Granite
7. ca. 2.1Ga Kenora-Fort Frances (Kabetogama) dike swarm
91°W
88°W
85°W
48°30'N
46°30'N
0 100 200 km
N
Nipigon Embayment
Nipigon sills
Lake Nipigon
BSF
Sibley Group
Thunder Bay
Osler Group
Slate Islands
Trans-Superior Tectonic Zone
Coldwell Alkalic Complex
Duluth Complex
Isle Royale
IRF
MR
NSF
Lake Superior
Thiel Fault
Ontario Michigan
MIF
Michipicoten Island
Kapuskasing Structural Zone
Sault Ste. Marie
KF
DF
Animikie Group
Duluth
GLTZ
Marquette Range Supergroup
Niagara Fault Zone
Penokean Orogen
Wolf River Batholith
Abbreviations
GLTZ - Great Lakes Tectonic Zone
NSF - North Shore Fault
MR - Midcontinent Rift
BSF - Black Sturgeon fault
IRF - Isle Royale Fault
KF - Keweenaw Fault
MIF - Michipicoten Island Fault
DF - Douglas Fault
Paleozoic
Illinois and Michigan basins
Mesoproterozoic
Midcontinent Rift
Clastic sedimentary rocks
Volcanic rocks (1.11-1.09 Ga)
Intrusive rocks (1.11-1.09 Ga)
Anorogenic/Post-orogenic suites
Sibley Group-sandstone, carbonate >1.54 Ga
Granitoid rocks (1.48-1.77 Ga)
Paleoproterozoic
Penokean Orogen and Related Rocks
Granitoid and volcanic arc rocks (1.89-1.84 Ga)
Animikie Group and Marquette Range Supergroup (ca. 2.1-1.85 Ga)
Archean
International boundary
Selected alkalic complexes (see Sage, this volume for more detail)
Dikes
Faults
Thrust faults

Figure 16.2. Major Proterozoic geological elements in the Lake Superior region. *Modified from* Hoffman (1989), Franklin et al. (1982) and Sutcliffe (1987).

an examination of the north shore of Lake Superior. In the latter part of the nineteenth century, copper was produced from the Mamainse Point and Michipicoten Island areas. Between 1871 and 1875, a number of traverses were made to establish a route for the Canadian Pacific Railway line. The route for the Canadian National Railways line to the north was explored between 1905 and 1908.

Wilson's (1910) comprehensive report and geological map of the Nipigon area was published in 1910, as Memoir 1 of the Geological Surveys Branch of the Canadian government. Early accounts with an emphasis on economic geology of the region include Tanton's (1931) report on the geology and mineral deposits of Proterozoic rocks in the Thunder Bay area and Thomson's (1954) report on the geology of the Mamainse Point copper area.

PALEOPROTEROZOIC ANIMIKIE GROUP

In Ontario, the Paleoproterozoic Animikie Group sedimentary rocks form a homoclinal sequence southwest of Thunder Bay (*see* Figure 16.2). The entire Animikie Basin occupies parts of Ontario, Minnesota, Wisconsin and Michigan and has been divided into 2 segments by the younger Midcontinent Rift (Morey 1983; *see* Figure 16.2). The northwest segment is further separated into 2 parts by the Duluth Complex. Different formational names have been applied east and west of the gabbro (Figure 16.3). The western part of the northern segment contains the Mesabi and Cuyuna iron ranges of Minnesota and the eastern part contains the Gunflint iron range of Ontario. The Mille Lacs Group, a basal sequence in the northwestern part of the Animikie Basin, is not present in Ontario. The southeast segment, which is correlated with the northwest segment on the basis of lithological similarity (Morey 1983), is occupied by the Marquette Range Supergroup (*see* Figure 16.3). This supergroup is much thicker and more diverse than the Mille Lacs and Animikie groups and is interrupted by several unconformities.

In the northern part of the Animikie Basin, in Minnesota and Ontario, Animikie Group rocks rest unconformably on Archean basement (Moorhouse 1960; Morey 1983) and Early Proterozoic dikes which have been dated by Rb-Sr methods at 2120±67 Ma (Southwick and Day 1983). South of the Great Lakes Tectonic Zone, the Animikie Group rocks were deformed during the Penokean Orogeny (Sims et al. 1980; Morey and VanSchmus 1988).

In Ontario, the Animikie Group is subdivided into the Gunflint Formation and the overlying Rove Formation (*see* Figure 16.3). The Gunflint Formation (Figure 16.4) consists of a lower unit up to 15 m thick of gravelly, basal conglomerate, ferruginous carbonate, chert, jasper, algal chert, hematite and magnetite taconite; a middle unit up to 35 m thick of argillite tuff, argillite, carbonate and chert; and an upper 80 m thick unit of thin-bedded chert and ferruginous carbonate, chert, jasper and silicate taconite (Moorhouse 1960). Abundant microfloral remains are present in the chert of the lower member (Barghoorn 1971; Edhorn 1973). The Gunflint Formation is a complex sedimentary assemblage with rapid vertical and horizontal changes (Goodwin 1960; Moorhouse 1960).

The variable mineralogy of the Gunflint Iron Formation (Moorhouse 1960; Floran and Papike 1975) includes silica minerals (chert, chalcedony and quartz), oxides (magnetite, hematite, goethite and limonite), carbonates (dolomite, ferroan dolomite, ankerite, siderite and minor calcite), silicates (greenalite, stilpnomelane, minnesotaite, chamosite and chlorite) and minor sulphide (pyrite). Both stromatolitic and taconitic facies of iron formation are present.

The Rove Formation consists of a lower part of black, locally pyritic shales which grades upward into shales interbedded with arkosic greywacke deposited by turbidity currents (Geul 1970, 1973). Drilling indicates that the Rove Formation is approximately 500 m thick in Pardee Township, in excess of 600 m in the vicinity of Squaw Bay on the Sibley Peninsula (Geul 1973) and approximately 1000 m thick in Minnesota (Morey 1967). The formation lacks graded beds, but is characterized by numerous directional features such as ripple marks, flute casts and rill marks. These features indicate currents flowed in a southeast direction perpendicular to the inferred shoreline (Morey 1967).

MESOPROTEROZOIC PRE–KEWEENAWAN ROCKS IN THE NIPIGON AREA

The Nipigon Embayment (*see* Figure 16.2), previously named the Nipigon Plate by Stockwell et al. (1972), consists predominantly of a sequence of Mesoproterozoic, pre–Keweenawan sedimentary rocks of the Sibley Group and the Keweenawan Nipigon diabase sills. These rocks overlie the Archean Superior Province and define a broad basinal structure extending north of Lake Superior for 160 km. In addition, pre–Keweenawan alkali granite and related volcanic rocks, and minor lamprophyre dikes are also present. Lithological units and tectonic events in the Nipigon area are summarized in Figure 16.1.

Pre–Keweenawan Igneous Rocks

Sutcliffe and Greenwood (1982) showed that a Proterozoic anorogenic magmatic event predated the Sibley Group sedimentary rocks in the northern part of Lake Nipigon. Rocks associated with this event are predominantly subvolcanic quartz-alkali feldspar porphyry to alkali granite, but include minor fragmental dacite to rhyolite, which are sparsely preserved beneath the diabase sills in the northern part of Lake Nipigon (Figures 16.5, 16.6, 16.7 and 16.8). The quartz-feldspar porphyry to alkali granite intrusion on English Bay of Lake Nipigon (*see* Figure 16.5) is dated at 1536.7^{+10}_{-23} Ma by the U-Pb method (Davis and Sutcliffe 1985). This age is comparable to, although slightly older than, a group of anorogenic plutons that were intruded in a wide belt from southern California to Labrador in the time interval 1.4 to 1.5 Ga (Silver et al. 1977; Anderson 1983). In addition, Davis and Sutcliffe (1985) reported a variety of Mesoproterozoic ages from felsic clasts in conglomerate at

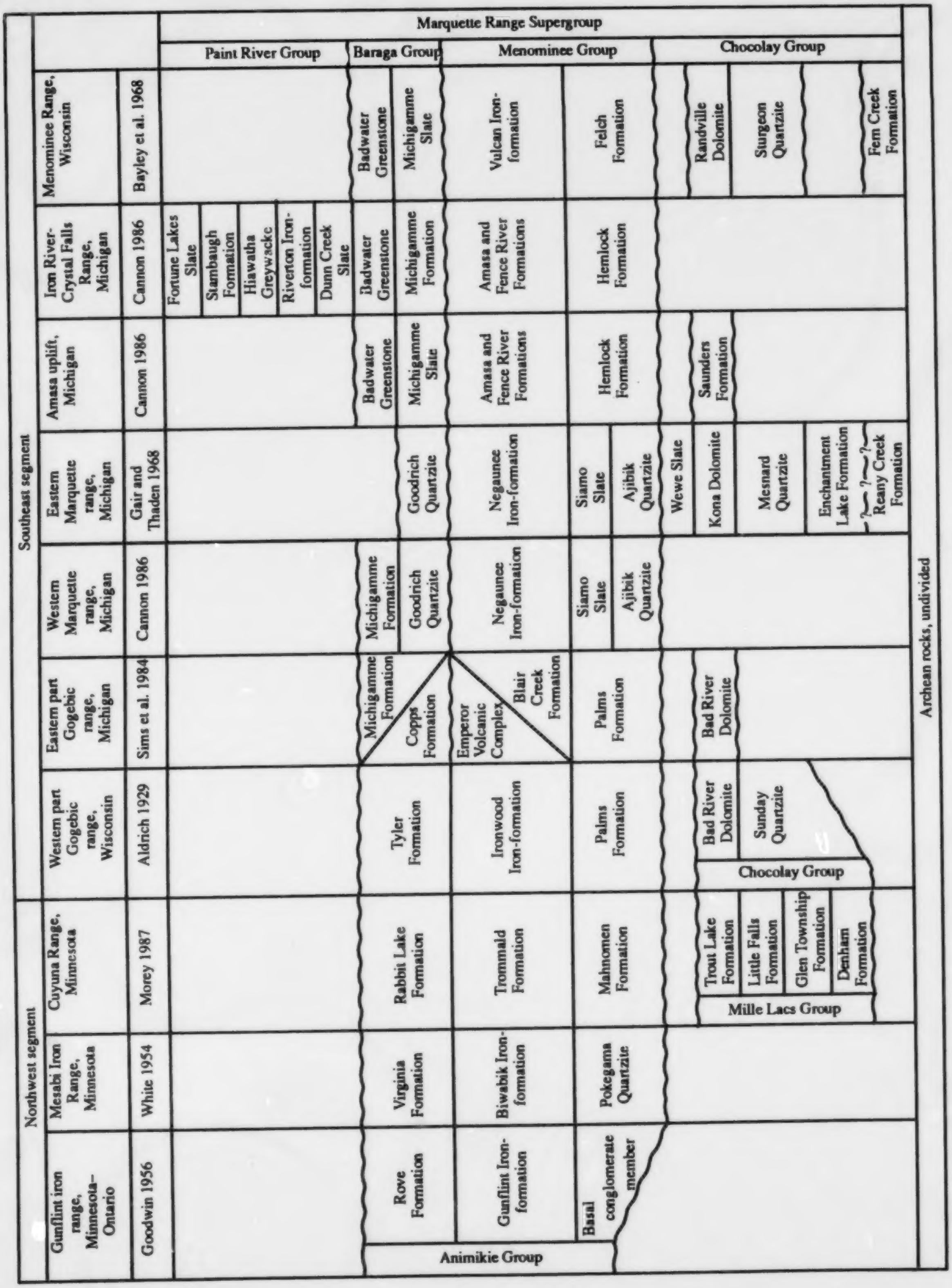

Northwest segment
Southeast segment
Gunflint iron range, Minnesota–Ontario
Goodwin 1956
Rove Formation
Gunflint Iron-formation
Basal conglomerate member
Animikie Group
Mesabi Iron Range, Minnesota
White 1954
Virginia Formation
Biwabik Iron-formation
Pokegama Quartzite
Cuyuna Range, Minnesota
Morey 1987
Rabbit Lake Formation
Trommald Formation
Mahnomen Formation
Trout Lake Formation
Little Falls Formation
Glen Township Formation
Denham Formation
Mille Lacs Group
Western part Gogebic range, Wisconsin
Aldrich 1929
Tyler Formation
Ironwood Iron-formation
Palms Formation
Bad River Dolomite
Sunday Quartzite
Chocolay Group
Eastern part Gogebic range, Michigan
Sims et al. 1984
Michigamme Formation
Copps Formation
Emperor Volcanic Complex
Blair Creek Formation
Palms Formation
Bad River Dolomite
Western Marquette range, Michigan
Cannon 1986
Michigamme Formation
Goodrich Quartzite
Negaunee Iron-formation
Siamo Slate
Ajibik Quartzite
Eastern Marquette range, Michigan
Gair and Thaden 1968
Goodrich Quartzite
Negaunee Iron-formation
Siamo Slate
Ajibik Quartzite
Wewe Slate
Kona Dolomite
Mesnard Quartzite
Enchantment Lake Formation
? ? ?
Reany Creek Formation
Amasa uplift, Michigan
Cannon 1986
Badwater Greenstone
Michigamme Slate
Amasa and Fence River Formations
Hemlock Formation
Saunders Formation
Iron River-Crystal Falls Range, Michigan
Cannon 1986
Fortune Lakes Slate
Stambaugh Formation
Hiawatha Greywacke
Riverton Iron-formation
Dunn Creek Slate
Badwater Greenstone
Michigamme Formation
Amasa and Fence River Formations
Hemlock Formation
Menominee Range, Wisconsin
Bayley et al. 1968
Badwater Greenstone
Michigamme Slate
Vulcan Iron-formation
Felch Formation
Randville Dolomite
Sturgeon Quartzite
Fern Creek Formation
Marquette Range Supergroup
Paint River Group
Baraga Group
Menominee Group
Chocolay Group
Archean rocks, undivided

Figure 16.3. Correlation chart for the Paleoproterozoic strata of the Animikie Basin (*modified from* Morey and Van Schmus 1988).

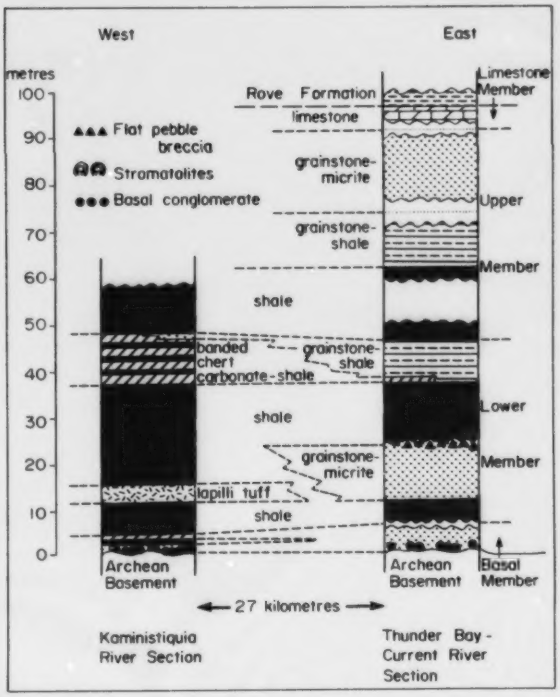

West
East
metres
100
90
80
70
60
50
40
30
20
10
0
▲▲▲ Flat pebble breccia
Stromatolites
●●● Basal conglomerate
Rove Formation
limestone
grainstone-micrite
grainstone-shale
shale
banded chert
grainstone-shale
carbonate-shale
shale
grainstone-micrite
lapilli tuff
shale
Limestone Member
Upper
Member
Lower
Member
Basal Member
Archean Basement
Archean Basement
27 kilometres
Kaministiquia River Section
Thunder Bay - Current River Section

Figure 16.4. Stratigraphy and lateral correlation of the Gunflint Formation from the Kaministiquia River to Current River (*modified from* Franklin et al. 1982).

the base of the Osler Group on Wilson Island. These clasts suggest that felsic magmatism also occurred at *circa* 1.6 and 1.7 Ga in the Lake Nipigon and northern Lake Superior area.

The granitic rocks exposed in northwestern Lake Nipigon contain moderately to strongly sericitized and hematitized orthoclase and oligoclase with quartz, chloritized biotite, zircon and fluorite. Quartz-feldspar porphyry and granite locally contain numerous inclusions of felsite, porphyry and pumice fragments. Volcanic rocks include debris flow, welded tuff and tuff breccia (Sutcliffe and Greenwood 1982).

The granite on English Bay has elevated abundances of zirconium, yttrium, and rare earth elements (REE), except for europium, which displays a pronounced negative anomaly. A typical analysis is given in Table 16.1. As reported by Davis and Sutcliffe (1985), the REE pattern is similar, although higher in total REE, to the more differentiated phases of the Wolf River Batholith, a 1.48 billion-year-old anorogenic batholith in Wisconsin (Anderson and Cullers 1978), and to 1.76 billion-year-old metaluminous granites and rhyolites in Wisconsin (Smith 1983). These rocks are interpreted as melts of crustal origin.

An arcuate, reverse fault, which follows the east shore of the northern part of Lake Nipigon (Sutcliffe and Greenwood 1982), provides evidence for a caldera structure in northern Lake Nipigon. This feature may have been associated with the emplacement of the English Bay Granite and associated volcanic rocks.

Figure 16.5. Mesoproterozoic quartz-feldspar porphyry from islands in English Bay of Lake Nipigon. The porphyry contains numerous inclusions of felsite and pumiceous rhyolite.

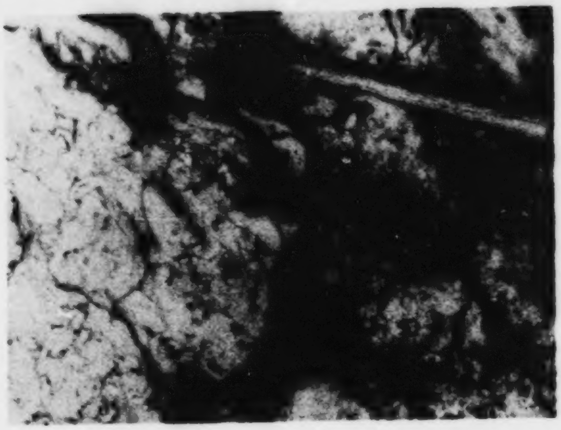


Figure 16.6. Mesoproterozoic volcanic conglomerate, south of Castle Bay, northeast shore of Lake Nipigon. The conglomerate contains subrounded clasts of felsite and clasts of amygdaloidal basalt with bleached rims.

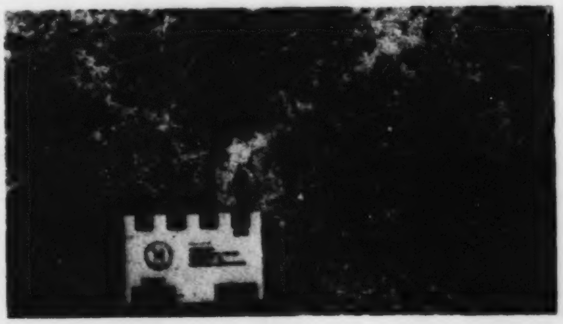


Figure 16.7. Mesoproterozoic volcanic breccia from the Mountain Islands in Lake Nipigon. Angular flow banded clasts of rhyolite are contained in an arkosic matrix. The photograph is taken of a large block which is not *in situ*.

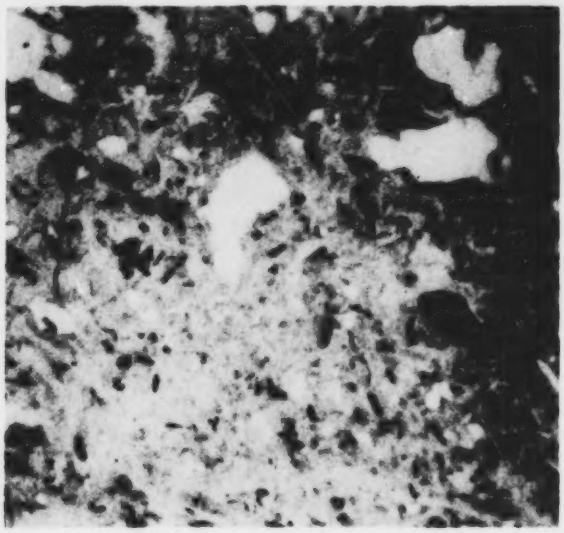


Figure 16.8. Photomicrograph of welded tuff from Humboldt Bay on the east shore of Lake Nipigon. The tuff forms a thin rind overlying outcrops of Archean rocks. Quartz and biotite phenocrysts in a matrix of devitrified glass shards are visible in the thin section. Plane light.

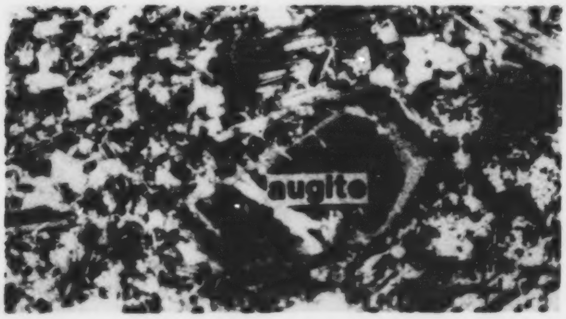


Figure 16.9. Photomicrograph of zoned augite phenocrysts in a camptonite dike. The dike is from Caribou Lake and contains xenoliths of garnet granulite from the lower crust. The matrix consists of plagioclase, sanadine, pyroxene, biotite, oxides and carbonate. Crossed polars, sample 80–337, *modified from* Sutcliffe (1988).

Numerous minor dikes of ultrabasic lamprophyre, camptonite and trachyte of presumed pre–Keweenawan age are present in the Caribou Lake area approximately 30 km northwest of Lake Nipigon (Sutcliffe 1988). The dikes, which are generally less than 1 m in width, are intrusive into Archean basement, but are not observed to intrude Mesoproterozoic diabase. The dike suite has alkalic characteristics and may be a result of the same magmatic event to that which produced the anorogenic granitoids at English Bay of Lake Nipigon.

The ultramafic lamprophyre dikes contain pseudomorphic olivine phenocrysts (replaced by serpentine and carbonate) and phlogopite phenocrysts in a matrix of phlogopite and carbonate with minor apatite and chromite. Camptonite (alkali basaltic lamprophyre) dikes (Figures 16.9 and 16.10) contain megacrysts of kaersutite, andesine, augite, anorthoclase and ilmenite in a matrix of plagioclase, augite, biotite, carbonate and locally sanadine. Apatite and relict olivine phenocrysts (Figures 16.11 and 16.12) are also locally present. Trachyte dikes predominantly contain hematitized alkali feldspar with minor sericite and chlorite. These dikes are geochemically characterized by high alkalis, normative nepheline and enriched light rare earth elements (LREE).

Sibley Group Sedimentary Rocks

The Sibley Group (Figures 16.13, 16.14, 16.15, 16.16 and 16.17; Coates 1972; Franklin et al. 1980) is a red bed sequence preserved in a fault-bounded basin elongated in a northerly direction and deepening to the east and south. Drilling has indicated the sequence has a maximum thickness of 420 m (Ojakangas and Morey 1982). The thickest part of the sequence is adjacent to the Black Sturgeon fault, a normal fault which forms the eastern boundary of an asymmetric graben extending north from Lake Superior to Lake Nipigon.

The lowest unit of the Sibley Group, the Pass Lake Formation, unconformably overlies Animikie Group sedimentary rocks near Pass Lake (Franklin et al. 1980) and Archean basement rock. The Pass Lake Formation consists of quartz arenite with locally derived conglomerate lenses. The sand is dominantly composed of quartz, but also contains feldspar and rock fragments. The Pass Lake Formation is overlain unconformably by the Rossport Formation, a red, fine-grained arenaceous to clayey dolomite unit. The clastic detritus becomes finer grained toward the top of the unit and a stromatolitic member is present in the middle part of the formation. The Rossport Formation is transitional into an upper unit of red shale and mudstone, the Kama Hill Formation, which contains sedimentary structures such as mud cracks indicative of shallow water and periodic dessication.

Small outcrop areas of quartz arenite in northern Lake Nipigon that overlie and are locally intercalated with volcanic rocks are characterized by metre-scale cross-beds and may have been deposited in an eolian environment (Sutcliffe and Greenwood 1982). These remnants may have been part of a sheet of quartz arenite that was continuous with the Pass Lake Formation of the Sibley Group to the south or alternatively may represent sedimentation in separate basins.

Franklin et al. (1980) suggested a marine origin for the Sibley Group; however, Ojakangas and Morey (1982) suggested evaporation of a freshwater lake as an alternative to a marine environment. This latter hypothesis is supported by Cheadle (1986), who suggested that the Sibley Group was deposited in an alluvial-playa lake environment analogous to the present interior basin of Australia.

A Rb-Sr whole rock isochron on dolomites and shales of the Sibley Group gives an age of 1339±33 Ma (Franklin et al. 1980), suggesting the sedimentary rocks are at least 200 Ma older than the volcanic rocks of the Keweenawan

Table 16.1. Representative analyses of Proterozoic igneous rocks from the Lake Superior area.

	1	2	3	4	5	6	7	8	9	10
SiO_2	49.8	51.2	49.6	68.8	42.2	47.9	49.29	49.50	48.35	74.87
TiO_2	2.50	1.28	1.62	0.57	0.78	2.73	.92	3.59	1.40	.20
Al_2O_3	10.3	14.8	15.6	13.60	2.96	11.70	15.39	13.34	15.15	12.63
Fe_2O_3	13.8	13.3	13.0	3.26	17.99	15.56	12.17	16.66	11.63	2.38
MgO	9.7	6.1	6.2	0.58	31.1	8.97	7.57	4.23	7.19	.35
MnO	.17	.19	.19	0.05	0.24	.21	.17	.19	.18	.03
CaO	9.9	10.4	10.4	2.18	4.16	9.48	11.77	7.34	11.31	1.04
Na_2O	2.8	2.2	2.7	2.84	.21	2.57	2.28	3.22	2.51	1.95
K_2O	.70	.46	.49	.4	.44	.67	.36	1.50	1.86	6.50
P_2O_5	.54	.18	.17	0.08	.06	.20	.07	.42	.42	.04
Rb	12.2	7.6	10.7	112	–	–	–	52	46	250
Sr	379	217	214	126	–	–	–	622	860	51
Y	23	24	28	27	6.90	28	26	36	21	82
Zr	226	128	154	355	13.4	–	79	267	166	525
Nb	16.9	9.3	9.6	19	–	–	–	24	48	113
Ta	1.00	0.57	0.60	–	–	–	–	ND	–	–
Th	1.8	3.0	1.7	12.6	–	–	–	ND	–	–
Ni	202	93	116	20	871	305	115	70	100	14
Cr	824	138	122	10	2549	430	243	33	267	14
La	24.5	15.3	13.2	96	9.80	25.0	5	34.0	52	180
Ce	60.9	33.5	31.4	195	17.0	62.0	11	78.0	105	345
Pr	8.6	4.1	4.4	21	2.20	7.50	–	9.80	–	–
Nd	39.4	16.7	19.3	75	9.80	34.0	6	44.0	40	120
Sm	8.7	4.2	5.1	12	2.20	7.70	3	10	8	31
Eu	2.63	1.34	1.65	2.30	.620	2.20	0.9	3.10	2.1	1.1
Cd	7.15	4.29	5.49	7.80	1.90	7.20	3	9.0	6.0	19
Tb	0.97	0.70	0.90	1.00	.280	1.10	–	1.30	–	–
Dy	5.22	4.32	5.62	5.30	1.50	5.70	2.9	7.70	4.0	20.0
Ho	0.93	0.94	1.15	1.00	.300	1.10	–	1.40	–	–
Er	2.33	2.68	3.21	2.50	.720	2.70	–	3.80	–	–
Tm	0.29	0.38	0.47	0.37	.090	.350	–	.480	–	–
Yb	1.78	2.47	3.00	2.30	.580	2.10	1.5	3.0	1.5	10.0
Lu	0.24	0.37	0.45	0.34	.080	.280	–	.460	–	–

Keweenawan Igneous Rocks

1. Average of 4 primitive olivine-augite basalt flows from lower suite of Osler Group (Lightfoot et al., in press).
2. Average of 4 primitive aphyric basalt flows from central suite of Osler Group (Lightfoot et al., in press).
3. Average of 4 primitive aphyric basalt flows from upper suite of Osler Group (Lightfoot et al., in press).
4. Quartz porphyry intruding Osler Group, St. Ignace Island (88–ARS–2027; Lightfoot et al., in press).
5. Peridote from Beardmore cone sheet, picritic suite, Lake Nipigon (81–273; Sutcliffe 1987).
6. Olivine-phyric picritic dike, chill, picritic suite, Lake Nipigon (83–251; Sutcliffe 1987).
7. Diabase sill, olivine tholeiite, Lake Nipigon (81–288; Sutcliffe 1987).
8. Diabase sill, quartz tholeiite, southwest of Thunder Bay (87–ARS–0161; Smith and Sutcliffe, unpublished data).

Pre–Keweenawan Igneous Rocks

9. Camptonite dike, Caribou Lake area (80–RHS–396; Sutcliffe 1986).
10. Alkali-granite, English Bay, Lake Nipigon (82–RMS–143; Sutcliffe 1986).

ND Not determined

Major element oxide data have been normalized to 100% on an anhydrous basis. Major elements in weight %, trace elements in ppm. All analyses by the Geoscience Laboratories Section, Ontario Geological Survey. Details of analytical methods are reported in references for each sample.

Supergroup. In northern Lake Nipigon, the porphyry intrusion on English Bay, dated at $1537^{+10}_{-2.3}$ Ma (Davis and Sutcliffe 1985), is associated with volcanic fragmental rocks that are intercalated with quartz arenite. This may provide a date for the initiation of Sibley Group sedimentation. The Sibley Group sedimentary rocks are intruded by diabase sills dated at 1109.7±2 Ma (Davis and Sutcliffe 1985).

MESOPROTEROZOIC KEWEENAWAN SUPERGROUP

Rocks of the Keweenawan Supergroup were deposited in and marginal to the Midcontinent Rift at approximately 1.1 Ga (Wallace 1981; VanSchmus et al. 1982). The rift is

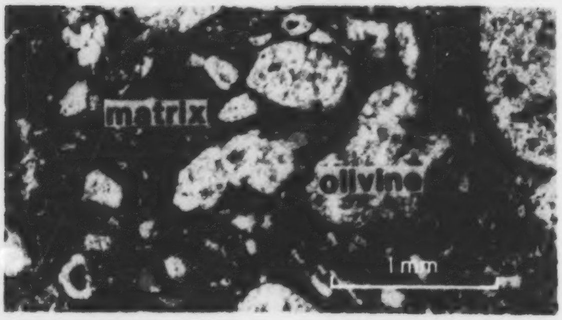

matrix
olivine
1 mm

Figure 16.10. Photomicrograph of olivine phenocrysts, pseudomorphed by serpentine and calcite, in a 0.5 m wide lamprophyre dike from Caribou Lake. The matrix consists of phlogopite, carbonate and opaques. Crossed polars, sample 80–1099, photograph *courtesy of* Sutcliffe (1988).

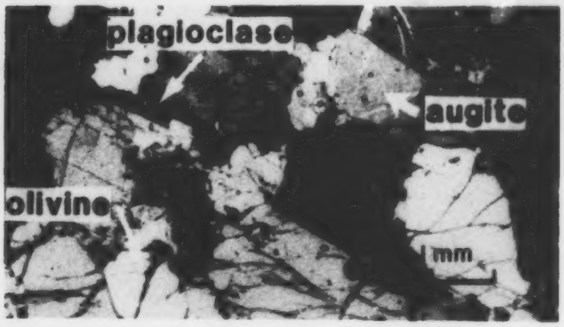

plagioclase
augite
olivine
1 mm

Figure 16.11. Photomicrograph of peridotite cumulate from the Beardmore ring dike on the east shore of Lake Nipigon. Olivine and clinopyroxene are cumulus grains and minor plagioclase is intercumulus. Crossed polars, sample 81–273.

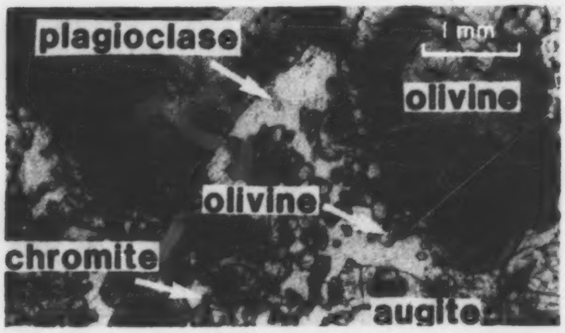

plagioclase
1 mm
olivine
olivine
chromite
augite

Figure 16.12. Photomicrograph of olivine gabbro from Jackfish Island, Lake Nipigon. Large olivine phenocrysts have zoning defined by opaque inclusions, small olivine grains are poikilitically enclosed by plagioclase. Augite is relatively minor. Euhedral chromite is enclosed by plagioclase. Plane light, sample 82–34.

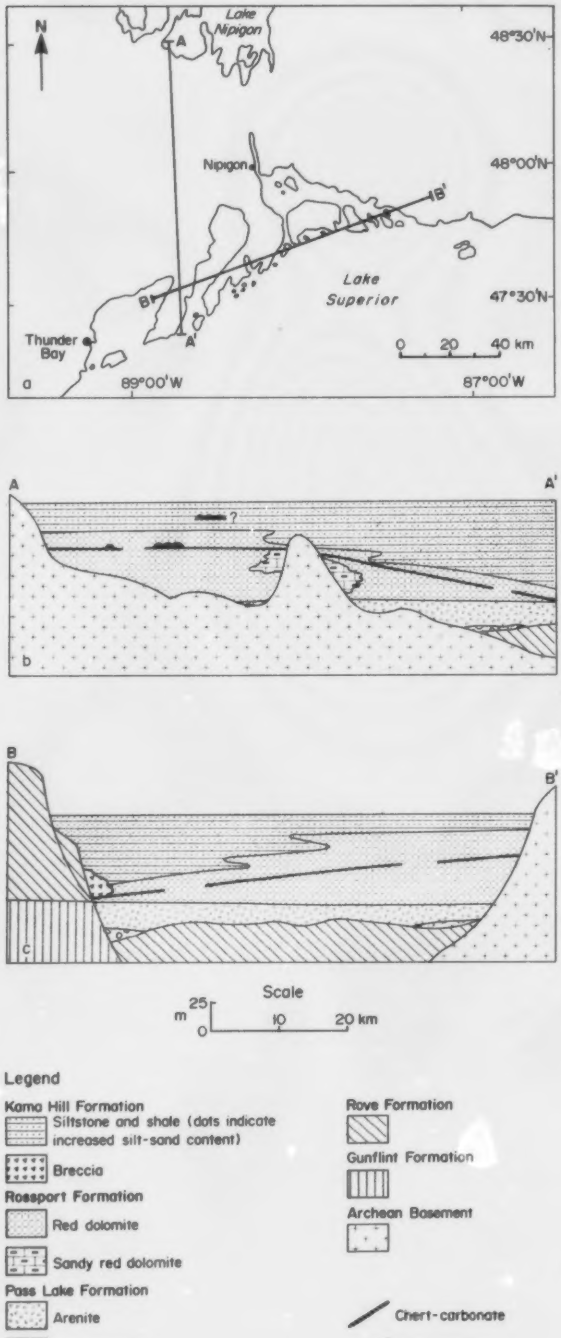

N
Lake Nipigon
A
48°30'N
Nipigon
48°00'N
B'
Lake Superior
B
47°30'N
A'
Thunder Bay
0 20 40 km
a
89°00'W
87°00'W
A
A'
?
b
B
B'
c
Scale
m 25
0
10
20 km
Legend
Kama Hill Formation
Siltstone and shale (dots indicate increased silt-sand content)
Breccia
Rossport Formation
Red dolomite
Sandy red dolomite
Pass Lake Formation
Arenite
Conglomerate
Rove Formation
Gunflint Formation
Archean Basement
Chert-carbonate
Stromatolite

Figure 16.13. Hypothetical regional cross sections of the Sibley Group (*modified from* Franklin et al. 1980): **a)** location of cross sections; **b)** north-south cross section; and **c)** east-west cross section.

only exposed in the Lake Superior area and is covered by Phanerozoic sedimentary rocks to the west and east. The main volcanism occurred between 1109 Ma (Davis and Sutcliffe 1985) and 1086 Ma (Palmer and Davis 1987; *see* Figure 16.1). The generalized stratigraphy of the supergroup is given in Table 16.2. A major paleomagnetic

Figure 16.14. Outcrop of Mesoproterozoic Sibley Group quartz arenite, probably the Rossport Formation, unconformably overlying Archean metasedimentary rocks which dip subvertically on Highway 11, west of Beardmore.

Figure 16.16. Mesoproterozoic stromatolitic dolomite of the Rossport Formation of the Sibley Group, west of Disraeli Lake. This block which is not *in situ* shows cross sections of the stromatolites (photograph *courtesy of* J. Scott).

Figure 16.15. Nipigon River south of Nipigon, showing a section of the Sibley Group (white) beneath a diabase sill (photograph *courtesy of* J. Scott).

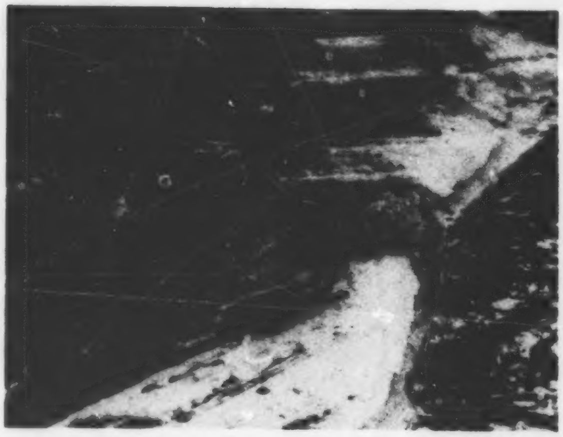


Figure 16.17. Cross-bedded quartz arenite of the Sibley Group on the north shore of Livingstone Point in northwestern Lake Nipigon. The sandstones consist of well-sorted quartz arenite and are exposed beneath a diabase sill.

reversal (Palmer 1970; Books 1972; Halls and Personen 1982) is an important tool for the correlation of Keweenawan sequences around Lake Superior. The reversal separates Lower Keweenawan rocks with a reversed magnetic polarity from Middle Keweenawan rocks with a normal polarity, although possible complications in this stratigraphy may exist at Mamainse Point (Massey 1979). In some cases, the polarity can be recognized on a basis of aeromagnetic data (e.g., Halls 1974) due to the strong remnant component. The Upper Keweenawan Supergroup consists of dominantly postvolcanic sedimentary rocks.

Keweenawan volcanic rocks consist predominantly of subaerial, tholeiitic basalt flows erupted mainly from fissures and to a lesser extent from central volcanoes (Green 1982; Figures 16.18, 16.19, 16.20, 16.21 and 16.22). Minor hyaloclastite units occur on Michipicoten Island (Annells 1974). Aluminum-rich olivine tholeiite is the most abundant compositional type, followed by transitional (weakly alkalic) iron-titanium-rich basalt, icelandite and rhyolite (Figures 16.23a and 16.23b; Green 1982). Green (1982) identified 8 basalt plateaus within the Lake Superior area, each forming areally extensive units with stratigraphic thicknesses of 1 to 6 km. Individual flows have lateral extents of up to 90 km (White 1960) and are typically a few metres thick with some flows greater than 100 m in thickness. Some of the major rhyolite units are traceable for up to 40 km and probably represent ash-flow eruptions (Green 1982).

Keweenawan interflow sedimentary rocks typically comprise 1 to 2% of the stratigraphic section and were deposited in fluviatile environments and locally lacustrine

Table 16.2. Stratigraphy[a] of the Keweenawan Supergroup (*modified from* Cannon et al. 1989).

Stratigraphic Correlation	Sedimentary Units	Volcanic Units	Intrusive Units
Paleozoic			
Upper Keweenawan	Bayfield Group and Jacobsville sandstone		
	Oronto Group (1094.1 Ma near base)		
Middle Keweenawan	Interflow sediments	Michipicoten Island (1085.5 Ma)[b] Portage Lake 1094.4 Ma[c] 1096.0 Ma Mamainse Point, North Shore Volcanic (upper), Osler Group (upper)	Duluth Complex[d], Mellen Gabbro
Magnetically normal			
	1097 Ma		
Magnetically reversed			Logan sills
Lower Keweenawan	Bessemer Quartzite Puckwunge Formation	North Shore Volcanics (lower), Powder Mill Group, Osler Group (lower) 1097.6[e] Ma 1107.5 Ma	Nipigon Sills (1108.8 Ma)[e], Baraga Dikes
Pre–Keweenawan	(Archean crystalline rocks, Paleoproterozoic and Mesoproterzoic metasedimentary and metavolcanic rocks)		

[a] Units listed within the Middle and Lower Keweenawan volcanic and intrusive units are at least in part correlative. Stratigraphic order is not implied by the order in which they are listed.

[b] Radiometric ages from Palmer and Davis (1987).

[c] Radiometric ages from Davis and Paces (1990).

[d] Includes some reversely polarized phases.

[e] Radiometric ages from Davis and Sutcliffe (1985).

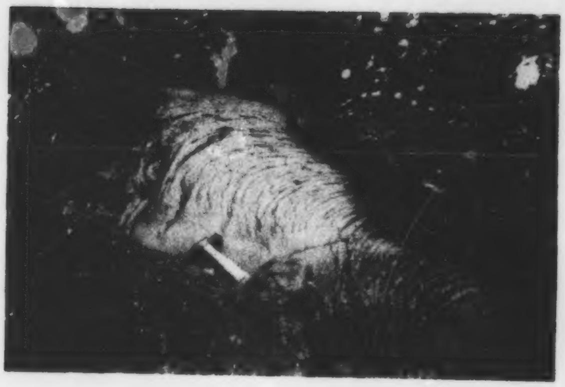


Figure 16.18. Ropy flow top on thin pahoehoe flow, overlain by the amygdaloidal base of another flow, from the lower part of the Osler Group on the north side of Wilson Island in Lake Superior.

Figure 16.19. Pipe amygdules in the basal part of a flow overlying conglomerate from the Osler Group on Puff Island, Lake Superior.

environments (Ojakangas and Morey 1982). Merk and Jirsa (1982) showed that paleocurrent patterns indicate that streams flowed towards the centre of the basin.

Figure 16.20. Polygonal fracturing in basalt flow from the South Shore Basalt Member of the Michipicoten Island Formation, Michipicoten Island, Lake Superior.

Figure 16.21. Glomeroporphyritic "daisy stone" flow of the Osler Group, Highway 17 at Mamainse Point.

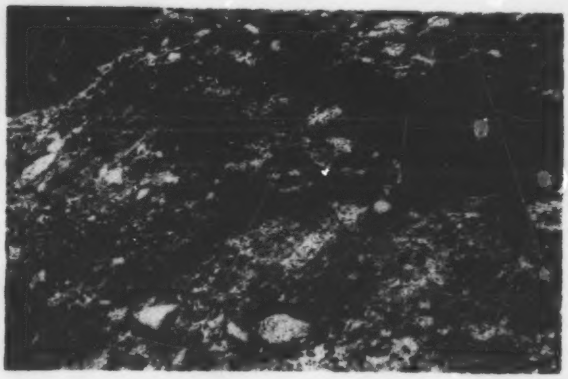


Figure 16.22. Conglomerate from the Osler Group on Puff Island, Lake Superior.

In Ontario, the volcanic rocks of the Keweenawan Supergroup are mostly exposed in the Black Bay Peninsula–St. Ignace Island area, Michipicoten Island and on Mamainse Point of Lake Superior. At these localities, rocks of the Keweenawan Supergroup dominantly consist of

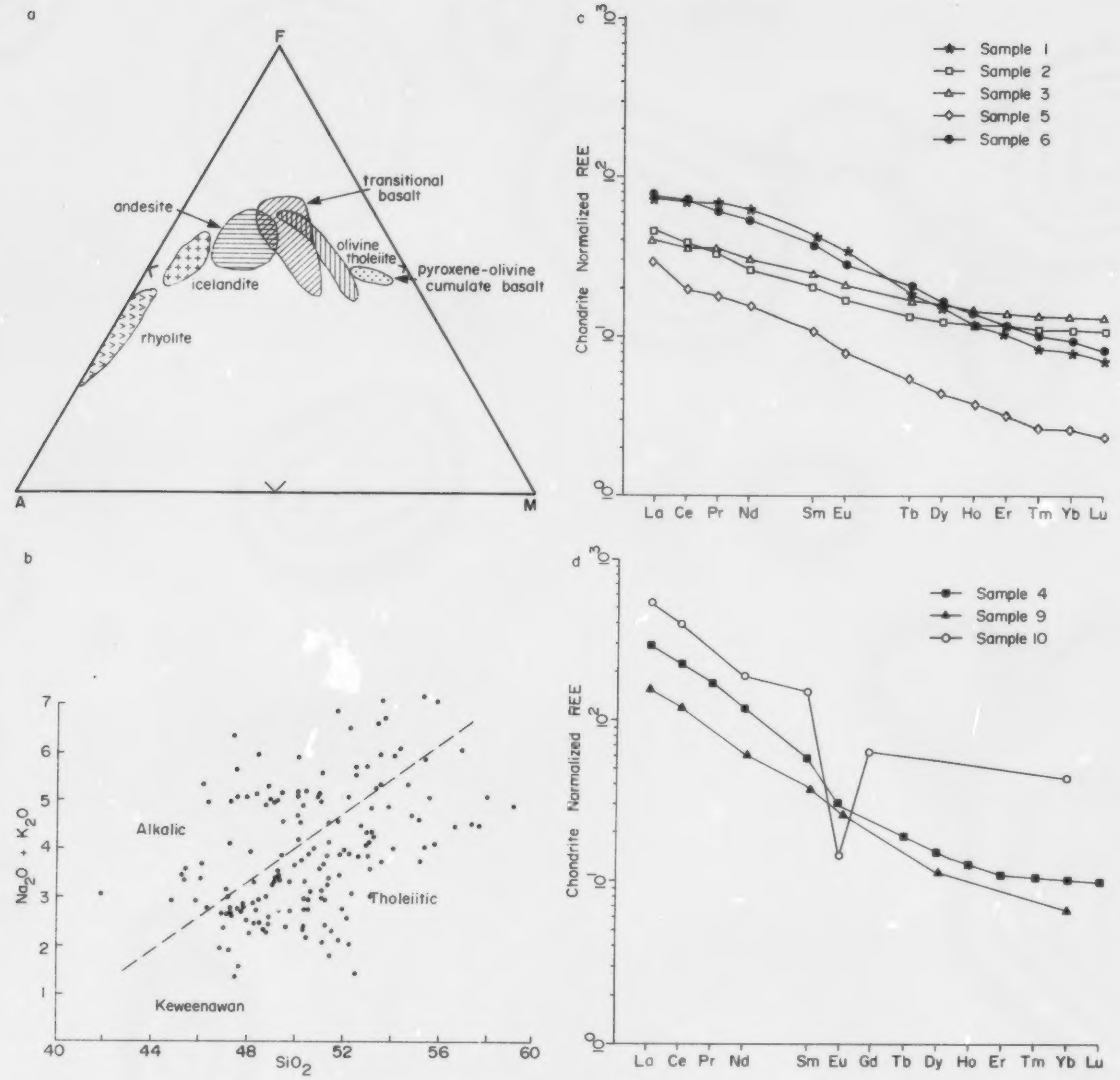


Figure 16.23. Compositional variation of Keweenawan basaltic rocks: **a)** AFM diagram for rocks of the North Shore Volcanic Group (*modified from* Green 1982); **b)** alkali/SiO_2 diagram of Keweenawan mafic and intermediate lavas with alkali basalt-tholeiitic basalt discriminant line (*modified from* Green 1982); **c)** and **d)** chondrite normalized REE plots for analyses in Table 16.1 (normalized values *modified from* Nakamura 1974).

subaerial tholeiitic basalts, with lesser amounts of andesite, rhyolite and interflow sedimentary rocks of largely fluvial origin (*see* Figures 16.18 to 16.22; McIlwaine and Wallace 1976; Annells 1973, 1974; Wallace 1981). Lesser accumulations of volcanic rocks are present on the Slate Islands (Sage 1983) and at other localities along the northeast shore of Lake Superior, such as Cape Gargantua (Ayres 1969). In Ontario, Upper Keweenawan postvolcanic sedimentary rocks are largely exposed in the vicinity of Sault Ste. Marie.

Metamorphic grade within the Keweenawan volcanic rocks of Ontario is very low. In the Osler Group of Black Bay Peninsula, McIlwaine and Wallace (1976) recognized a lower zone of prehnite-pumpellyite facies and an upper zone of zeolite facies which characterizes most of the area. On Michipicoten Island, metamorphism is of very low grade. Annells (1974) recognized zeolite facies assemblages in the Mamainse Point Formation which forms the stratigraphically lowest parts of the island. Michipicoten Island Formation basalts and andesites, however, display an unaltered glassy mesostasis (Annells 1974; Palmer et al. 1988).

Black Bay Peninsula–St. Ignace Island

Osler Group tholeiitic flood basalts underlie most of St. Ignace Island and the Black Bay Peninsula area. These rocks have previously been described by Giguere (1975) and McIlwaine and Wallace (1976). Most of the sequence consists of magnetically reversed Lower Keweenawan flows, although a small section of magnetically normal Middle Keweenawan rocks is preserved near the top of the sequence (Halls 1974). The base of the section is marked by a quartz-feldspar-phyric rhyolite that overlies Sibley Group sedimentary rocks. McIlwaine and Wallace (1976) considered that the rhyolite is probably a series of flows, but may be a high-level intrusion. Conglomerate, locally containing abundant porphyry clasts, crops out sporadically along the north shore of the peninsula and adjacent islands and occurs both above and below the basal rhyolite. Quartz-feldspar-phyric rhyolite flows also occur in the Osler Group near the top of the magnetically reversed sequence and are particularly well exposed near Agate Point. The exposed thickness of the Osler Group in the Black Bay Peninsula area is approximately 2800 m.

Davis and Sutcliffe (1985) reported U-Pb ages from this sequence. The rhyolite from the base of the Osler Group is 1107.5^{+4}_{-2} million-years-old, whereas the rhyolite from Agate Point, below the magnetic reversal, is 1097.6±3.7 million-years-old. In addition, a variety of zircon ages from felsic clasts in the conglomerate at the base of the Osler Group suggests pre–Keweenawan felsic magmatism took place at about 1730 and 1600 Ma in the area.

The basalts consist of massive to amygdaloidal flows, with locally developed flow tops, pahoehoe structures and pipe amygdules (*see* Figures 16.18 and 16.19). The flows are typically aphyric with well-developed ophitic textures in massive parts of the flows (*see* Figure 16.21). Amygdules are filled with calcite, agate, zeolites, prehnite, chlorite and quartz. Flow contacts and intercalated sedimentary rocks indicate that the regional dip of the sequence is approximately 6°S.

Interflow sedimentary rocks are most abundant in the lower part of the Osler Group and consist of conglomerate with lenses of cross-bedded sandstone and siltstone. An approximately 10 m thick conglomerate with pebble- to boulder-size basalt and felsic porphyry in a sandstone matrix is also present in the vicinity of the magnetic reversal near the top of the section. McIlwaine and Wallace (1976) also reported lenses of fine-grained volcanic detritus and cherty beds in the upper part of the Osler Group.

Sutcliffe and Smith (1988) identified a composite stratovolcano that overlies flood basalts of the Osler Group on St. Ignace Island. The volcanic rocks of the sequence, here named the St. Ignace Formation, consist of intercalated plagioclase glomerophyric basaltic rocks and quartz-feldspar-phyric rhyolite. These volcanic rocks are interpreted to be the extrusive equivalent of the underlying gabbro to anorthosite ring dike and central quartz-feldspar porphyry stock. The felsic rocks range from massive and flow-banded rhyolite to rhyolite breccia and welded tuff. The rhyolite commonly contains numerous inclusions of glomerophyric basalt, gabbro and andesite porphyry. Heterolithic and monolithic volcanic conglomerate and associated sandstone lenses are interbedded with the flows.

Michipicoten Island

Michipicoten Island, studied by Annells (1974), consists of a 4300 m thick sequence of basalt, andesite and rhyolite flows, volcaniclastic rocks with minor intercalated sandstone and conglomerate. Included in this estimate of stratigraphic thickness are concordant intrusions of fine-grained andesite, felsite and quartz porphyry. The sequence is unusual among the Keweenawan volcanic piles and rift-related sequences, in general, because of the abundance of intermediate and felsic rocks. The volcanic rocks have a southerly dip which increases from 45° to 50° on the north side of the island to 10° to 15° in the south. Neither the top nor the base of the sequence is exposed on the island. Geophysical evidence indicates that an east-trending fault with downthrows of 1 to 2 km to the north occurs along the north shore of the island (Hinze et al. 1966; Halls and West 1971).

Annells (1974) subdivided the flows into 2 formations and the following description is based on this work. The Mamainse Point Formation occupies the lowest part of the succession and consists of olivine tholeiites with intercalated conglomerates which have been intruded by quartz porphyry and felsite of the first intrusive phase. In the second intrusive phase, all of these rocks were intruded by andesite which locally differentiated to form granophyre. The Michipicoten Island Formation, which overlies the previous units, comprises thick flows of glassy andesite, quartz-normative basalts, rhyolite and associated volcaniclastic rocks and minor sedimentary rocks.

Palmer and Davis (1987) determined that the quartz porphyry of the first intrusive phase has a U-Pb zircon age of $1086.5^{+1.3}_{-3.0}$ Ma. This porphyry sample contained inherited zircon with an age of 1403±50 Ma, suggesting that the porphyry may have been derived from an older felsic source which may have been an anorogenic pluton. The Michipicoten Island porphyry is approximately 22 my younger than the age of the Nipigon sills. This age difference probably spans the bulk of Keweenawan igneous activity.

Mamainse Point

The Mamainse Point Formation (Giblin 1969; Annells 1973; Massey 1983) is a 4300 to 6000 m thick sequence of subaerial tholeiitic flood basalts, with interbedded coarse polymictic conglomerate and is intruded by rhyolite and felsite. The sequence strikes north to northwest and dips at 15° to 45° west towards Lake Superior. Paleomagnetic studies by Palmer (1970) revealed that most of the sequence shows normal polarization, but 2 sequences of flows that show reverse polarization are present in the lower part of the section. Palmer (1970) suggested that the original magnetic sequence of reverse polarity followed by normal polarity, which is characteristic of other Keweenawan volcanic sequences, has been repeated by faulting in the Mamainse Point area. Subsequent geochemical studies by Massey (1979, 1983), however, indicate that the 2 reverse sequences are geochemically distinct and that 3 changes in magnetic polarity occurred.

Annells (1973) reported that the the Mamainse Point section contains over 300 basalt flows which typically range in thickness from 1.5 to 30 m. Flow textures include lower vesicular zones commonly containing pipe vesicles, upper vesicular zones with ropy pahoehoe surfaces, scoriaceous tops, and locally tumuli or small pressure ridges. Flow textures include medium-grained ophitic and diabasic, fine-grained intersertal, and plagioclase-phyric to glomerophyric flows.

Felsic intrusions in the Mamainse Point area include quartz porphyry, felsite and flow-banded rhyolite, and range in form from small plugs and dikes to thin intrusive sheets (Giblin 1969; Annells 1973).

Geochemistry of Keweenawan Basalts

Studies of Keweenawan basalts in Michigan and Minnesota indicate that the dominant basalt types include high-alumina olivine tholeiite (characterized by low potassium, phosphorus, titanium and high aluminum and enriched LREE) and transitional or weakly alkalic, iron-titanium basalts (higher potassium, phosphorus, titanium, iron, REE and lower aluminum; *see* Figure 16.23b, Green 1982; *see also* Figures 16.23c and 16.23d). Typical analyses of Keweenawan volcanic rocks are shown in Table 16.1.

Berg and Klewin (1988) reported that high-MgO lavas, with up to 23% MgO, occur in the basal part of the Mamainse Point sequence. The high-MgO lavas, which were interpreted by Berg and Klewin (1988) as representing picritic liquids, are intercalated with subordinate high-TiO_2 ferro basalts. The latter include the plagioclase glomeroporphyritic "daisy stone" flows. The high-MgO lava sequence is overlain by aphyric and ophitic tholeiitic basalts, some of which show petrographic evidence of crustal contamination (Massey 1980).

Geochemical studies of the Michipicoten Island rocks were reported by Annells (1974). Olivine tholeiites of the Mamainse Point Formation are characterized by high Al_2O_3 (14.8 to 18.3 weight %), high MgO (6.4 to 8.4 weight %), moderate TiO_2 (1.3 to 2.3 weight %) and low K_2O (0.3 to 0.6 weight %). Annells (1974) considered these basalts to be very similar to Keweenawan olivine tholeiites of the Keweenaw Peninsula, Minnesota and Mamainse Point. Michipicoten Island Formation andesites have considerably higher TiO_2, Zr and Ba than typical tholeiitic and calc-alkalic andesites. Annells (1974) suggested that high Ba and Zr abundances may be due to crustal contamination.

A study of the geochemistry of the Osler Group tholeiitic basalts, based on a section along Nipigon Straits, reveals 3 basalt suites (Figure 16.24; Lightfoot et al., in press). The suites are sufficiently distinct to preclude fault repetition of stratigraphy. The Lower Suite (lower 750 m of basalts) consists of augite-olivine-phyric flows with relatively high Mg-numbers (0.58 to 0.66), SiO_2 (48.2 to 53.5 weight %), TiO_2 (1.49 to 3.19 weight %), Gd/Yb (3.2 to 4.5) and by relatively low La/Sm (2.0 to 3.5). The Central and Upper Suite flows are aphyric or plagioclase-phyric and have lower Mg-numbers (0.62 to 0.30). Upwards through the Central Suite, the SiO_2 (51 to 48.5 weight %), La/Sm (2.5 to 6.0) and Gd/Yb (2.5 to 1.5) values decline.

Distinctions in the 3 suites of the Osler Group are illustrated in Figure 16.24 (*from* Lightfoot et al., in press). This data indicates that the Lower Suite magmas have been derived from higher degrees of melting of a garnet lherzolite source than the other suites (Lightfoot et al., in press). Secondly, variation in the Central and Upper suites is consistent with contamination by crustal material, as represented by felsic flows from St. Ignace Island, which are located at a stratigraphic interval between the Central and Upper suites. Evidence for crustal melting is also demonstrated by the U-Pb zircon data for felsic flows at Agate Point, which are stratigraphically equivalent to the top of the Central Suite, and have an inherited Archean zircon component (Davis and Sutcliffe 1985).

Postvolcanic Keweenawan Sedimentary Rocks

Sedimentary rocks form the upper part of the Keweenawan section and consist primarily of a red bed sequence that was deposited during regional subsidence following the cessation of volcanism. These rocks comprise conglomerate, shale and sandstones of the Oronto Group, the base of which is intercalated with the 1094 million-year-old Portage Lake Volcanics (Daniels 1982; Davis and Paces 1990) and unconformably overlying sandstones of the Jacobsville Group. Based on studies south of Lake Superior, the Oronto Group (*see* summaries by Daniels 1982; Ojakangas and Morey 1982) is subdivided into: the Copper Harbour Conglomerate, an alluvial fan-fluvial clastic wedge; the Nonesuch Shale, a grey reduced cupriferous, pyritiferous, carbonaceous shale, probably deposited in a lacustrine environment; and the Freda Sandstone, a fluvial red bed sandstone and mudstone sequence. The sandstone of the Jacobsville Group, which appears to unconformably overlie the Oronto Group (Cannon et al. 1989), is a sequence of feldspathic to quartzose fluvial sediments (Ojakangas and Morey 1982). The sandstone of the Jacobsville Group is found in Michigan and eastern Lake Superior and is probably correlative with the Bayfield Group in the western Lake Superior area (Ojakangas and Morey 1982). The unfossiliferous sandstone of the Jacobsville Group is assigned to the late Keweenawan, based largely on paleomagnetic evidence that the sandstone is only slightly younger than the Oronto Group (Roy and Robertson 1978). Based on paleomagnetic studies, the correlative Bayfield Group is considered to have been affected by the late Keweenawan compressive event (Watts 1981).

In Ontario, outcrops of these groups occur in the vicinity of the eastern end of Lake Superior. Beneath eastern Lake Superior, seismic profiles indicate that these rocks have a maximum thickness of approximately 8 km (Cannon et al. 1989).

A 60 m thick section of folded sandstone crops out in the Alona Bay–Mica Bay area near Mamainse Point and has been correlated with the Freda Formation by Hamblin (1961), Giblin (1969) and Annells (1973). The lava flows in

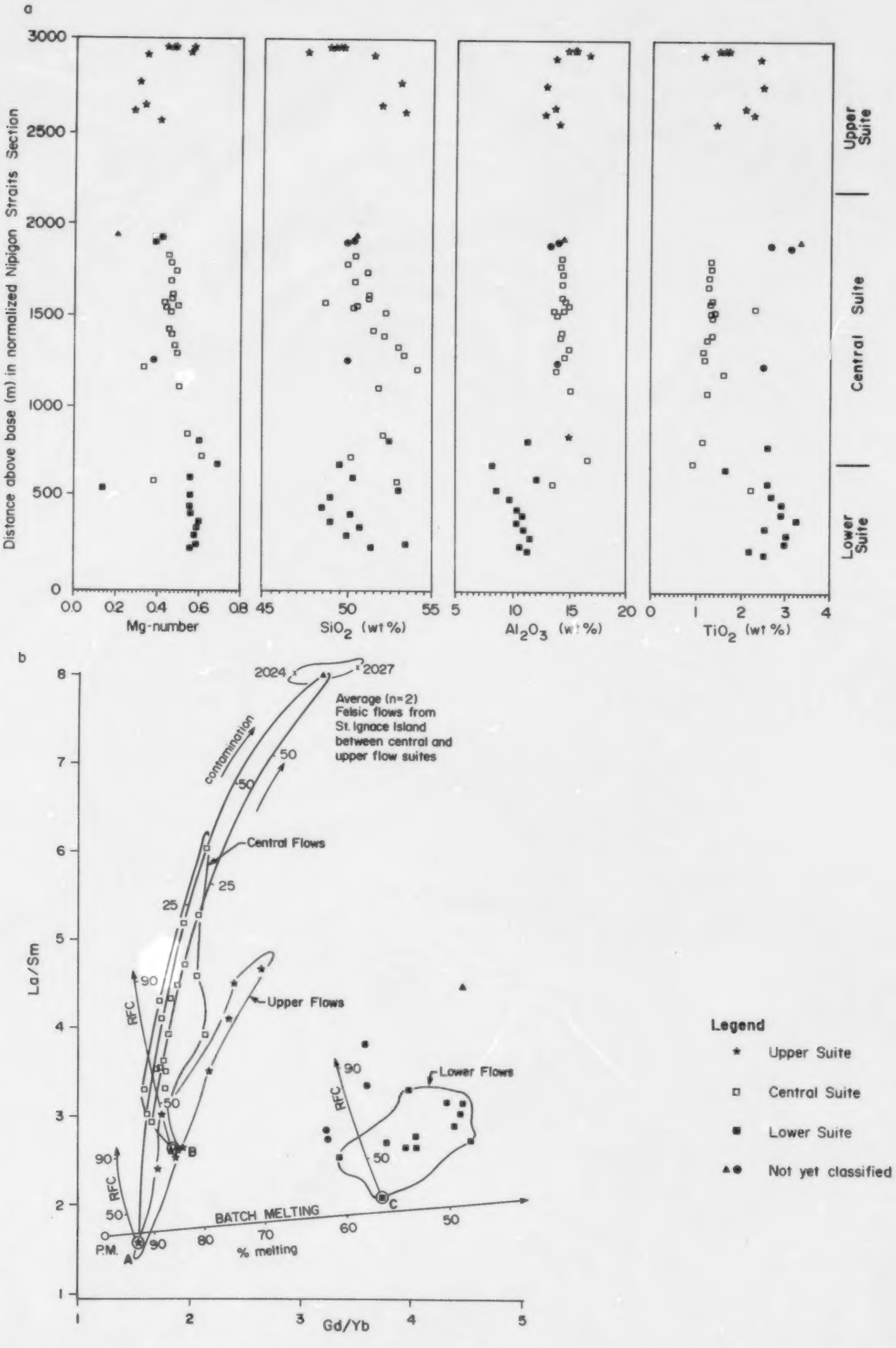


Figure 16.24. Compositional variation of the Osler Group flood basalt sequence on Black Bay Peninsula (*from* Lightfoot et al., in press): **a)** variation in Mg-number, SiO_2, Al_2O_3 and TiO_2 versus stratigraphic height showing division of the Osler Group into 3 suites; and **b)** variation in La/Sm versus Gd/Yb. Variations are modelled in terms of batch melting, Rayleigh fractional crystallization (RFC) and mixing with a felsic crustal melt.

the Mamainse Point–Alona Bay area were tilted and eroded before deposition of the Freda Formation, the base of which is a thin pebble conglomerate with fragments of Keweenawan basalt (Annells 1973). Ojakangas and Morey (1982) considered that the sandstone was largely derived from extrabasinal source rocks and was deposited by fluvial processes.

Outcrops of horizontally bedded quartzose sandstone of the Jacobsville Group occur in Ontario near Sault Ste. Marie. Ojakangas and Morey (1982) reported that transport directions for the sandstone of the Jacobsville Group, including measurements from Ontario, are toward the centre of the Lake Superior basin and are compatible with fluvial deposition.

MESOPROTEROZOIC KEWEENAWAN INTRUSIVE ROCKS

Keweenawan intrusive rocks are represented by a diverse range in composition and form that includes: hypabyssal tholeiitic diabase dike swarms and sills; layered tholeiitic mafic intrusions; central mafic to felsic ring complexes; and alkalic complexes. Weiblen (1982) reviewed characteristics of Keweenawan intrusions including the voluminous Duluth Complex, which outcrops in Minnesota. The Duluth Complex contains an early anorthositic series intruded by a later troctolite series, which typically consists of discrete layered plutons. Table 16.1 presents geochemical data for typical Keweenawan intrusive rocks in Ontario.

Thunder Bay Area

The Mesoproterozoic Logan diabase sills intrude sedimentary rocks of the Rove Formation in the area south and west of Thunder Bay (Pye 1953; Blackadar 1956; Geul 1970, 1973) and in northeastern Minnesota (Jones 1984; Weiblen et al. 1972). The sills and sheets dip gently to the southwest and form the caps of large cuestas. Smith and Sutcliffe (1987) identified 6 major sills in the Thunder Bay area with thicknesses of up to 45 m. The sills are characterized by a reverse magnetic polarity (Personen 1979) and, therefore, are probably of Lower Keweenawan age and were intruded at an early stage of rifting, comparable to the Nipigon sills in the Lake Nipigon area. The sills are intruded by olivine-tholeiite diabase dikes of the Pigeon River dike swarm (Geul 1970, 1973).

Typical sill sections (Figures 16.25a and 16.25b) are characterized by a narrow, centimetre thick, basal chill zone, which grades upward into ophitic and medium-grained plagioclase-phyric diabase (Smith and Sutcliffe 1987). The second sill from the base of the sequence is predominantly composed of red hematitized granophyre, but has a lower chilled zone of diabase. The mineralogy of the sills is dominantly labradorite, intersertal to ophitic augite with lesser pigeonite, and iron-titanium oxides. Micrographic quartz-alkali feldspar intergrowth is abundant in the upper part of some sill sections.

Geochemical studies of sills in the Thunder Bay area by the author (unpublished data) indicate that the Logan sills are quartz-normative tholeiites. The sills have similar geochemical characteristics to the iron-titanium basalt suite of Green (1982) and to the type Logan sills described by Weiblen et al. (1972). These characteristics include high TiO_2, FeO, K_2O, P_2O_5 and REE and low Al_2O_3 and MgO relative to the olivine tholeiite suite. Blackadar (1956) considered the granophyric zones to be the products of crustal assimilation. The presence of sills composed predominantly of granophyre in the Thunder Bay area, however, indicates that the granophyre could not be a result of *in situ* assimilation.

The Crystal Lake Gabbro is a Keweenawan layered intrusion that is intrusive into sedimentary rocks of the Animikie Group. Geul (1970) determined that the intrusion has 2 major limbs and documented the geology of the copper, nickel and PGE mineralization of the Great Lakes Nickel deposit which occupies the northern limb. The gabbro also intrudes the northeast-trending diabase dikes of the Pigeon River dike swarm (Smith and Sutcliffe 1987). Smith and Sutcliffe (1987) subdivided the northern limb into 4 major lithological zones defined on a basis of variations of cumulus phases and textures within the north limb of the intrusion (Figure 16.26). The Basal Zone (7 m) consists of fine- grained and aphanitic gabbro with inclusions of sediment. The Lower Zone (60 m) is characterized by heterogenous gabbro to anothositic gabbro with abundant pegmatitic patches and contains up to 10% disseminated chalcopyrite, pentlandite and pyrrhotite. The Lower Zone is overlain by

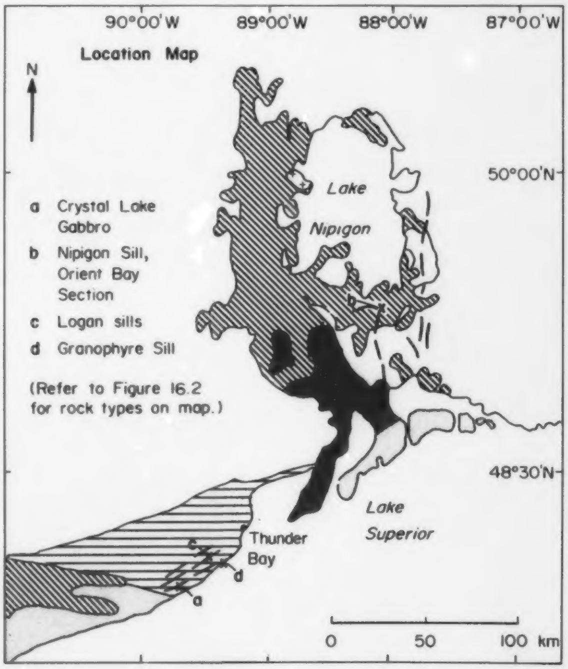


Figure 16.25a. Modal composition of sections of Keweenawan intrusive rocks in the Thunder Bay and Nipigon areas.

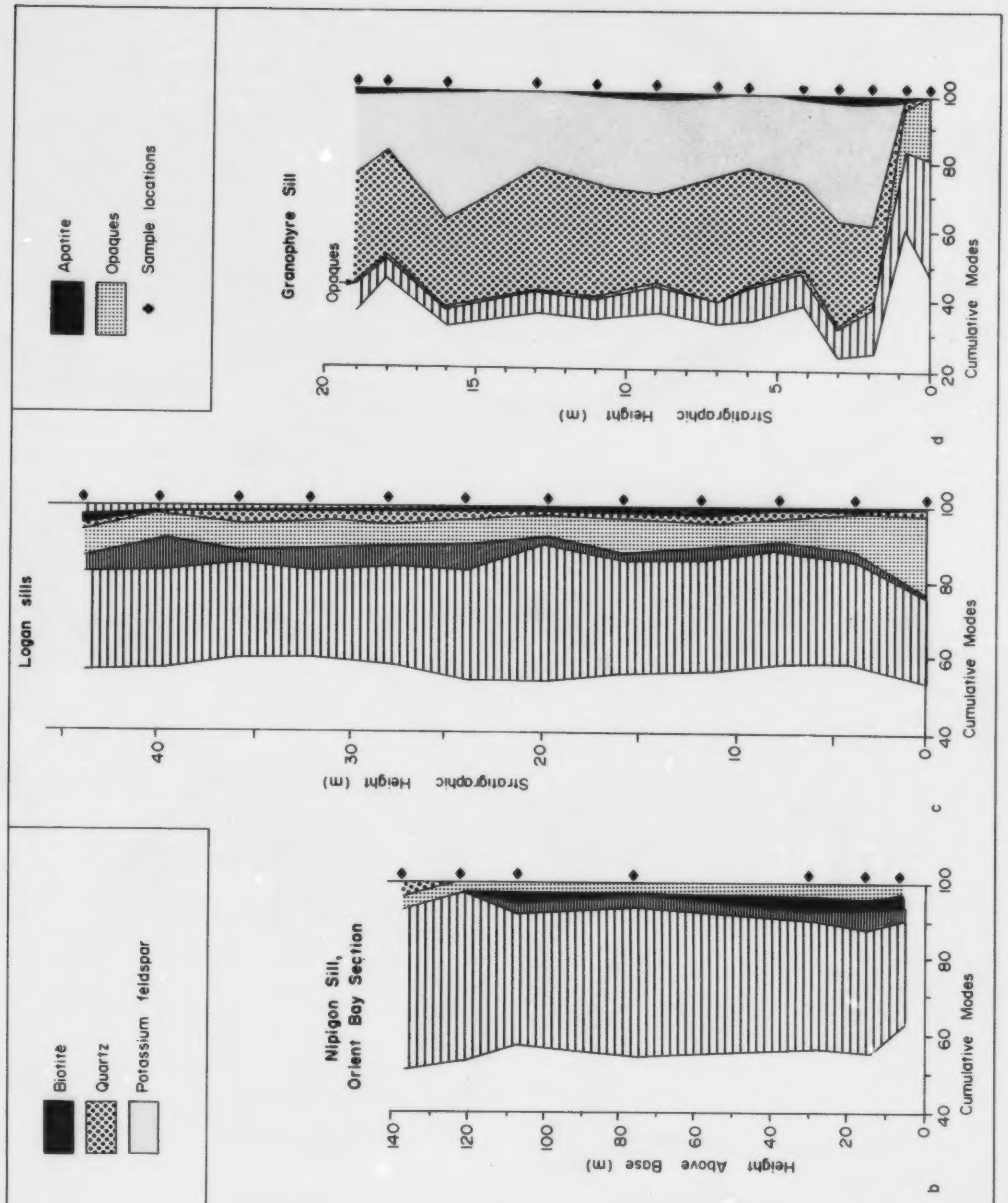
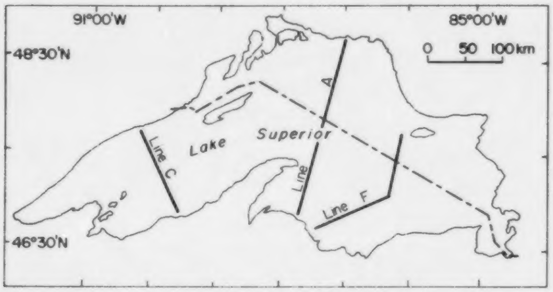


Figure 16.25b. Modal composition of sections of Keweenawan intrusive rocks in the Thunder Bay and Nipigon areas.

the Layered Zone (25 m), which contains distinct modal layers of anorthosite, gabbro and chromite-bearing anorthosite. The Upper Zone consists of medium-grained troctolite to olivine gabbro with wispy to intermittent modal layering.

The Pine River–Mount Mollie Intrusion is a composite, northeast-trending dike, 60 to 350 m wide, which extends for approximately 35 km from Crooks Township to McKellar Island in Lake Superior (Geul 1970; Smith and Sutcliffe 1987). The intrusion varies in composition from troctolite and gabbro through diorite to granophyre. The gabbroic rocks locally have steeply dipping modal layers and contain patches of pegmatitic gabbro. Sulphide mineralization occurs along the margins of the intrusion. The granophyre phase contains miarolitic cavities and locally breccia zones. The breccia zones consist of angular sedimentary fragments within a granophyre matrix, providing evidence for explosive emplacement of the phase. The intrusion is not contiguous at surface with the Crystal Lake Gabbro, but the common trend and presence of some comparable rock types suggest that the 2 intrusions may be contemporaneous.

On St. Ignace Island, gabbroic rocks of the St. Ignace Complex form a central ring complex and smaller sills which are intrusive into the volcanic rocks. These rocks were initially mapped by Giguere (1975) and reinvestigated by Sutcliffe and Smith (1988). The intrusive rocks range in composition from gabbro to anorthosite and consist of plagioclase, clinopyroxene, olivine and opaques. These rocks are coarse grained and exhibit igneous lamination and locally wispy, modal layering. A central felsic porphyry stock with an area of approximately 22 km² is contained within the gabbro ring dike. The stock is composed of massive red to grey quartz-alkali feldspar porphyry and contains abundant fine- to medium-grained plagioclase-phyric mafic inclusions near the contact with the gabbro ring dike. Textures in both the intrusive and volcanic rocks of the St. Ignace Island complex indicate that mafic and felsic magmatism were contemporaneous and that localized mixing of magmas occurred. Rounded chilled mafic inclusions, hybrid (compositionally intermediate) rocks and basaltic fragments in felsic welded tuff support this hypothesis. The gabbro is intrusive into reversely polarized Osler Group



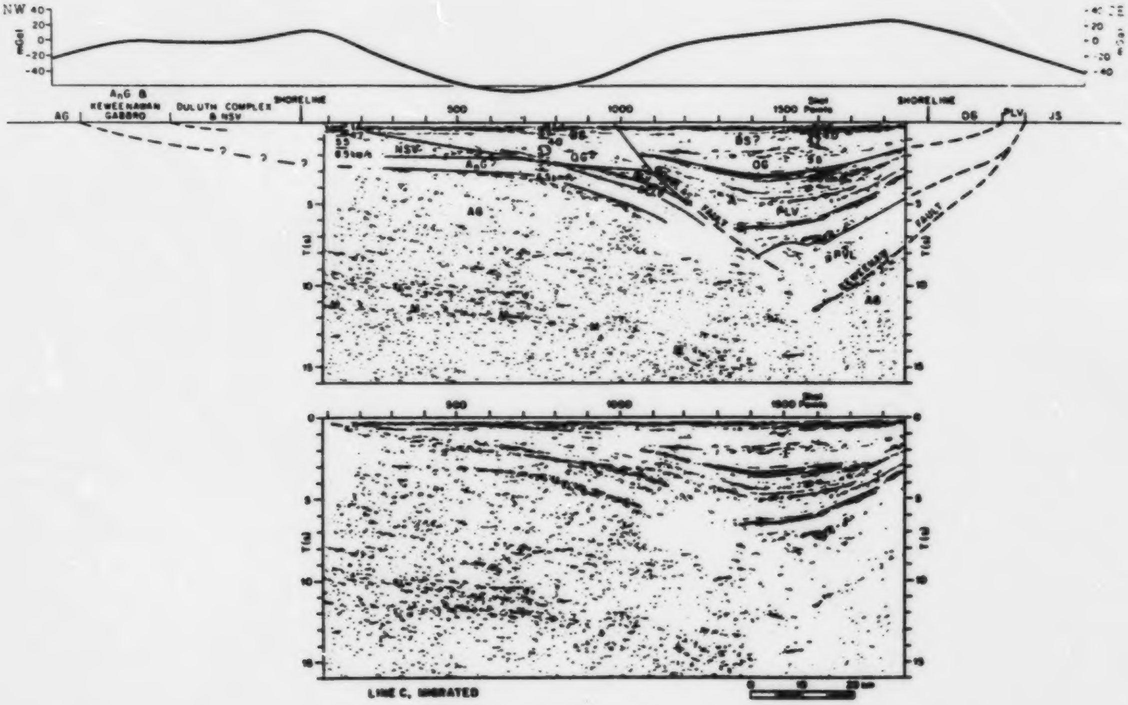
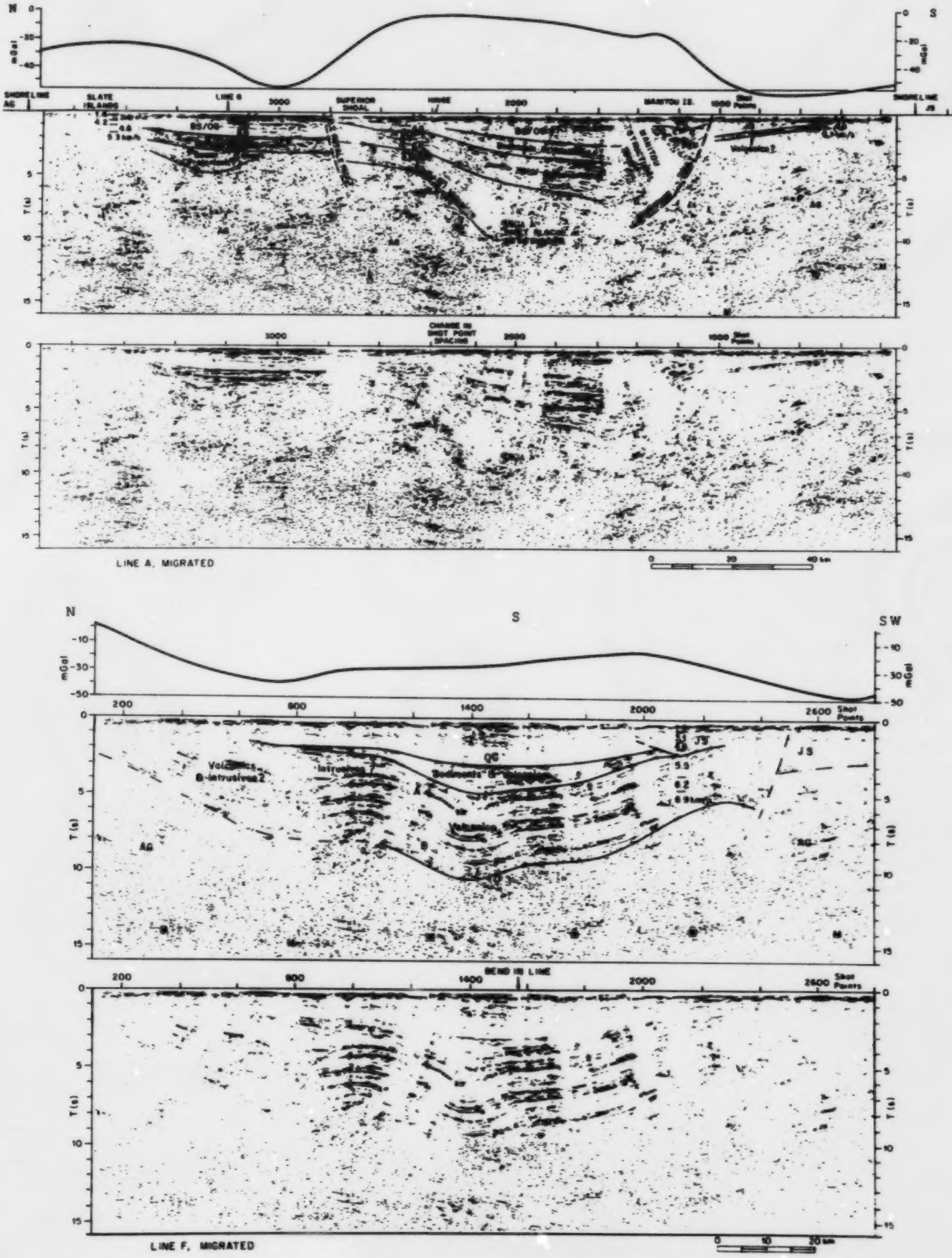


Figure 16.26. Seismic reflection profiles from the GLIMPCE experiment (Cannon et al. 1989): **a)** location map of seismic lines; **b)** Line C; **c)** Line A; and **d)** Line F. Top part of each diagram shows Bouguer gravity profile and interpreted seismic reflection profile. Bottom part of each diagram shows migrated seismic reflection profile.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S
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Volcanics?
T (s)
5
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CHANGE IN SHOT POINT SPACING
LINE A, MIGRATED
0
20
40 km
N
S
SW
-10
-30
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800
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Volcanics & intrusives?
Intrusives?
Sediments &
Volcanics
AG
M
BEND IN LINE
LINE F, MIGRATED
0
10
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Figure 16.26. Seismic reflection profiles from the GLIMPCE experiment (Cannon et al. 1989): **a)** location map of seismic lines; **b)** Line C; **c)** Line A; and **d)** Line F. Top part of each diagram shows Bouguer gravity profile and interpreted seismic reflection profile. Bottom part of each diagram shows migrated seismic reflection profile.

volcanic rocks. The felsic volcanic rocks associated with the complex are probably correlative with the reversely polarized Agate Point Rhyolite dated at 1098 Ma (Davis and Sutcliffe 1985).

The Moss Lake Intrusion on Black Bay Peninsula is a circular funnel-shaped body approximately 11 km in diameter, which ranges in composition from gabbro and gabbronorite to anorthositic gabbro (McIlwaine and Wallace 1976). The intrusion has a chilled contact adjacent to the volcanic rocks and contains a central body of basalt and quartz-feldspar porphyry which is interpreted to be a roof pendant. A petrographic study by Keeler (1971) indicated that the gabbro contains clinopyroxene, subordinate orthopyroxene, plagioclase (An_{55} to An_{70}), with locally developed quartz and quartz-alkali feldspar intergrowths. Olivine (Fo_{30}), ferroaugite, titanomagnetite and plagioclase (An_{45}) occur in the most fractionated rocks toward the centre of the intrusion.

The St. Ignace Island Complex and the Moss Island Intrusion have similarities to the central complexes of the British Tertiary Volcanic Province, such as the Mull and Ardnamurchan complexes (Thompson 1982). Features of the Keweenawan complexes which are similar to the British complexes include: a rift-related tectonic setting; the presence of volcanic rocks underlain by cogenetic central intrusions; coexisting basaltic and felsic magmas with evidence of hybridization; and emplacement of magmas controlled by ring fractures.

Lake Nipigon Area

Mafic igneous rocks of the Nipigon area consist of an early, volumetrically minor suite of picritic intrusions and a later suite of olivine tholeiite diabase dikes, sheets and sills (Figures 16.27 to 16.34; Sutcliffe 1987). These rocks are intrusive into the Sibley Group sedimentary rocks and Archean basement. The sills have a reversed magnetic polarity characteristic of the lower Keweenawan Supergroup (DuBois 1962). Zircon dating of the diabase sills indicates that they were intruded at 1108.8^{+4}_{-2} Ma and are contemporaneous or slightly older than Osler Group basaltic magmatism associated with the Midcontinent Rift (Davis and Sutcliffe 1985). To the south in the Nipigon Bay area, the sills dip under Osler Group flood basalts. This relationship in conjunction with the geochronology suggests that the sills may have contributed to crustal loading facilitating later eruption of volcanic rocks (Davis and Sutcliffe 1985; Sutcliffe 1987). Except for minor granophyre dikes, felsic rocks associated with the diabase are absent.

Petrographic descriptions and petrological characteristics of the mafic igneous rocks (*see* Figures 16.9 to 16.12, 16.35 and 16.36), are largely summarized from Sutcliffe (1987, 1989). One major diabase sill approximately 150 to 200 m thick is exposed over the Nipigon area and locally is overlain by erosional remnants of an upper sill of comparable thickness. Major sills have a textural stratigraphy consisting of: a lower chilled and aphanitic zone (2 to 3 m); an ophitic zone (30 to 60 m); a medium-grained zone with locally developed modal layering (40 to 70 m); a pegmatitic zone (20 to 40 m); and an upper chilled aphanitic zone (2 to 3 m). This stratigraphy (*see* Figure 16.25) indicates that, in most cases, the sills represent single cooling units.

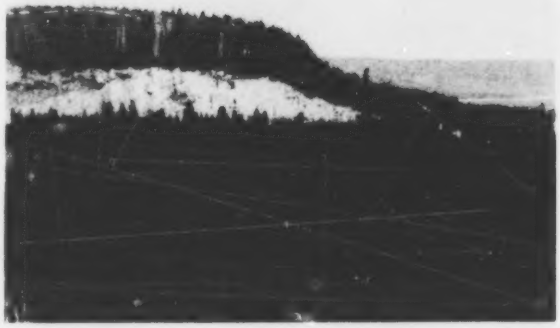

Figure 16.27. Typical exposure of section of Nipigon diabase sill on Caribou Island east of Thunder Bay. Diabase exposed at the top of the cliff shows weakly developed columnar jointing. Extensive diabase talus forms the base of the cliff, commonly obscuring the lower contact of the sill with the Rove Formation (photograph *courtesy of* J. Scott).

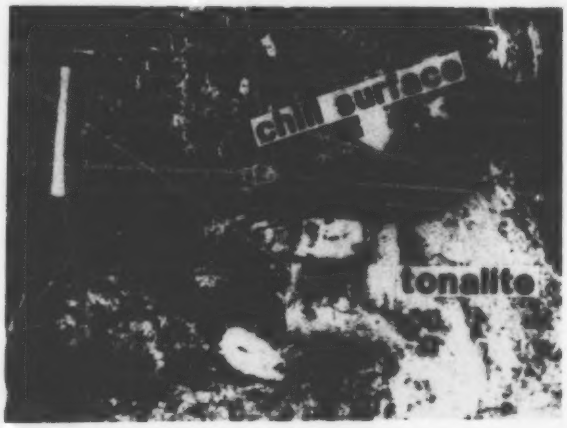

Figure 16.28. Typical exposure of the lower contact of a diabase sheet on the south shore of Berry Lake, west of the town of Armstrong. Lower contact of the sheet is polygonally fractured and dips at approximately 30°. Tonalite country rock is visible in the lower part of the photograph.

The intrusion of the diabase into the impure dolostones of the Rossport Formation of the Sibley Group has resulted in the development of contact metamorphic calc-silicate assemblages in the sedimentary rocks. Sutcliffe (1986) reported that within approximately 10 m of the contact with the diabase, the normally red weathered Rossport Formation is bleached white. Commonly developed calc-silicate assemblages are calcite + tremolite + forsterite and at higher grade, calcite + diopside + forsterite (Sutcliffe 1986). Contact zones of diabase with the sedimentary rocks reveal the diabase sheets propogated as fingers that subsequently coalesced to form sheets (*see* Figures 16.33 and 16.34).

Figure 16.29. Diabase chill on the upper contact of a sill showing polygonal fractures, from the west shore of Sturgeon Lake, southwest of Lake Nipigon. Fractures are defined by epidote alteration (light coloured) with an indistinct outer zone of amphibole.

Figure 16.30. Chill zone from the upper contact of a diabase sill with round miarolitic cavities lined with amphibole (dark) and wollastonite (white). Eagle Egg Islands in central Lake Nipigon.

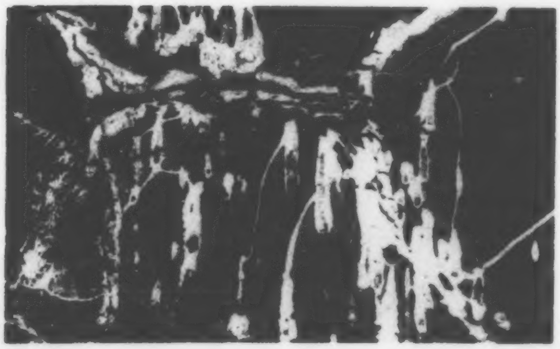

Figure 16.31. Elongated miarolitic cavities lined by epidote alteration in the upper chill zone of a diabase sill, Dawson Island, Lake Nipigon.

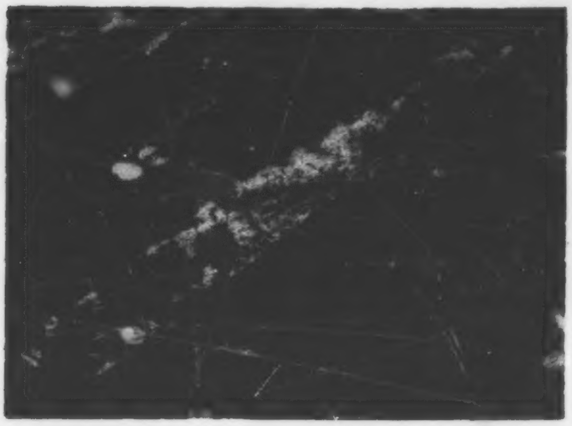

Figure 16.32. Cusp-shaped grooves in the upper surface of diabase on the east shore of Lake Nipigon, south of Wabinosh Bay. The grooves locally contain deformed remnants of deformed Sibley Group sedimentary rocks and are interpreted to result from the coalescing of diabase fingers. Note: Brunton compass for scale.

The major silicate mineral phases in the sills and dikes are plagioclase, variable amounts of augite and pigeonite, olivine, minor orthopyroxene and micrographic quartz-alkali feldspar intergrowths. The diabase sills crystallized from evolved liquids, characterized by low TiO_2, P_2O_5 and K_2O, similar in composition to the high Al_2O_3 olivine tholeiite flood basalts of the Midcontinent Rift. Compositional variation in the sills reflects cotectic crystallization of plagioclase with lesser olivine and pyroxene and movement of residual liquids enriched in H_2O, Na and Fe towards the top of the sections.

The Nipigon sills are fed by cone sheets and dikes. Northwest of Lake Nipigon, a large cone sheet 30 to 40 km in diameter intrudes the Archean rocks and is associated with a broad 30 mgal, positive, Bouguer gravity anomaly. East of Lake Nipigon centred near Beardmore, a second cone sheet system of comparable dimensions consists of diabase and peridotite. In general, dikes are not abundant in the Nipigon area, however, south of Lake Nipigon, the Fox Mountain dike, a composite diabase-gabbro intrusion, is emplaced into the Black Sturgeon graben and is also probably a feeder for diabase sills.

Picritic rocks in the Lake Nipigon area occur as dikes, ring dikes, small plugs and sills and have been described by Sutcliffe (1987). These rocks are present in the Jackfish Island area of northwest Lake Nipigon, the southeast shore of Lake Nipigon and in the Disraeli and Leckie lakes area, south of Lake Nipigon (*see* Figures 16.9 to 16.12, 16.35 and 16.36). The picritic rocks contain cumulus euhedral olivine and lesser clinopyroxene, with interstitial to poikilitic orthopyroxene and plagioclase (*see* Figure 16.12). The absence of aphyric chills and the ubiquitious cumulate textures of these rocks indicate that the intrusions were formed by accumulation of olivine at shallow crustal depths. The picritic rocks have higher TiO_2, P_2O_5 and K_2O than the

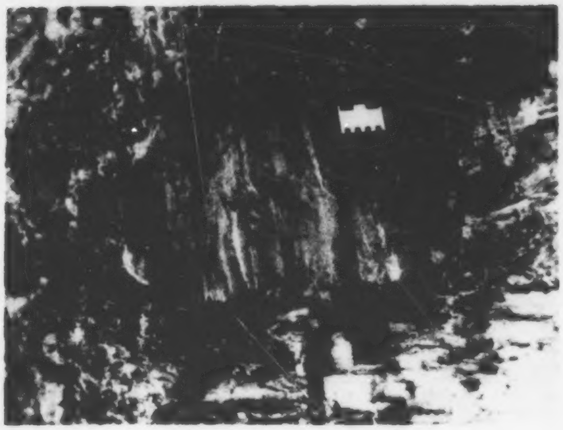

Figure 16.33. Cross section of a diabase finger intrusive into shale of the Rove Formation near the Great Lakes Nickel deposit, southwest of Thunder Bay.

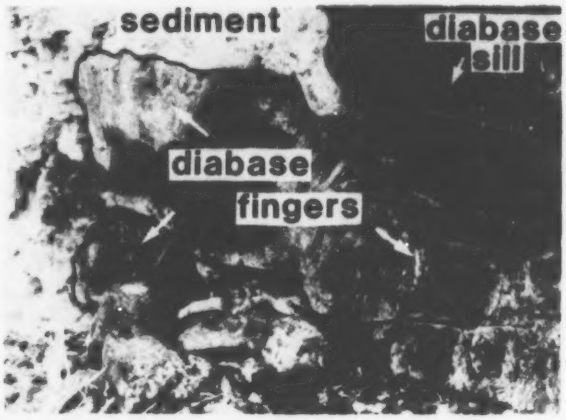

Figure 16.34. Diabase fingers intrusive into sedimentary rocks of the Sibley Group at Cooke Point, Lake Nipigon. Several fingers have coalesced and have wedges of deformed sediment between them.

diabases and are similar to, but more primitive than, iron-titanium basalts in the Keweenawan Supergroup.

Carbonatite and Alkalic Complexes

Carbonatite and alkalic complexes of Keweenawan age are emplaced along several zones of regional faulting within the Superior Province (*see* Sage, this volume). The most significant of these zones includes the Kapuskasing Structural Zone. During the Middle Proterozoic, the Kapuskasing Structural Zone localized the Lackner Lake, Shenango, Nemegosenda, Firesand and Clay–Howells alkalic complexes and the northern extension of the Thiel fault along which the Port Coldwell alkalic complex, Killala Lake Alkalic Complex and Prairie Lake Carbonatite Complex are intruded. Other intrusions of probable Keweenawan age are emplaced into faults extending northeast of Wawa.

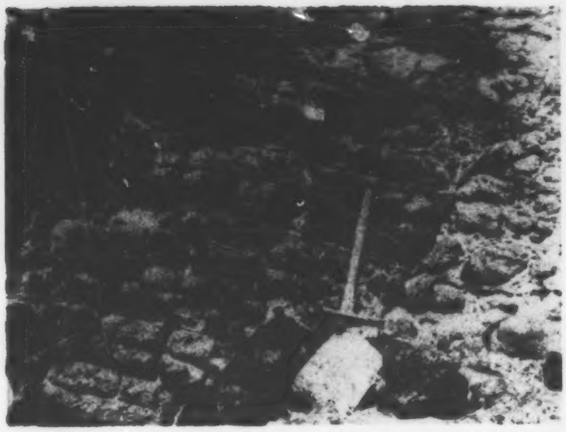

Figure 16.35. Outcrop of peridotite on the east shore of Lake Nipigon, south of Lion Bay in Kitto Township. Serpentinized fractures result in characteristic blocky weathering.

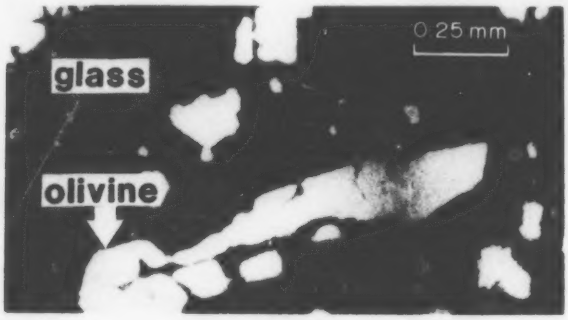

Figure 16.36. Photomicrograph of resorbed skeletal olivine in picritic chill of dike on English Bay, Lake Nipigon. Plane light, sample 83–251.

The Port Coldwell alkalic complex is the largest Mesoproterozoic alkalic complex (Puskas 1967) and consists of gabbroic to syenitic rocks which occur in 3 coalescing ring-shaped intrusive centres (Mitchell and Platt 1977). The Port Coldwell alkalic complex is considered in more detail by Sage (this volume), but is noted here because the 1108±1 Ma U-Pb zircon age for the complex (Heaman and Machado 1987) indicates that a major component of Keweenawan alkalic magmatism occurred at the same time as the tholeiitic magmatism such as that represented by the Nipigon sills.

TECTONIC SYNTHESIS

Paleoproterozoic Animikie Group and Penokean Orogen

In the southern part of the Animikie Basin, the Animikie Group sediments were deformed by the Penokean Orogeny. The deformed southern zone is separated from a virtually unmetamorphosed and undeformed northern zone (which includes the sequence in Ontario) by a tectonic front corresponding approximately to the Great Lakes Tectonic Zone (Morey and VanSchmus 1988; *see* Figure 16.2). The Great Lakes Tectonic Zone is defined as the boundary between Archean gneissic and greenstone-granitoid terranes (Sims et al. 1980). During the Penokean Orogeny, juvenile Paleoproterozoic crust was accreted onto the southern margin of the Superior Province (Barovich et al. 1989). The Niagara Fault Zone (*see* Figure 16.2) forms the boundary between the Superior Province and the overlying deformed Marquette Range Supergroup to the north and the accreted Wisconsin magmatic terrane to the south.

Hoffman (1989) considered that the Animikie Group is a synorogenic foredeep sequence that oversteps the rifted-passive margin sequence represented by the lower part of the Marquette Range Supergroup. Based on Sm-Nd studies of the Michigamme Formation in the upper part of the Marquette Range Supergroup, Barovich et al. (1989) have shown that the foredeep sediments were derived from the Paleoproterozoic arc terrane in Wisconsin. Deposition of the upper Michigamme Formation is contemporaneous with the initiation of suturing of the Wisconsin terrane to the craton margin and the onset of the Penokean Orogen at approximately 1.86 Ga (Hoffman 1989).

Carbonatite and alkalic intrusions were emplaced *circa* 1.9 Ga in several areas of the Superior Province (*see* Sage, this volume). In addition to being approximately coincident with the accretionary Penokean Orogen, this age is also coincident with several other major Paleoproterozoic events, including the Trans–Hudson Orogen (*see* Hoffman 1989).

Mesoproterozoic Pre–Keweenawan Events

The anorogenic granite at English Bay is of comparable age to anorogenic granitoids emplaced at 1.4 to 1.5 Ga in a wide belt extending from southern California to Labrador (Silver et al. 1977; Anderson 1983). Windley (1989) noted that these rocks are similar to the Phanerozoic anorogenic suite in the Hoggar–Niger–Nigeria region, which is associated with extensional tectonism prior to continental separation. Windley (1989) attributed the Proterozoic magmatic belt to pre–Grenvillian lithospheric stretching.

The time interval between Sibley Group sedimentation and Keweenawan magmatism suggests that the Sibley Group is not related to Keweenawan rifting. It has been suggested that Sibley Group sedimentation may have been controlled by the graben structure bounded by the Black Sturgeon fault (Card et al. 1972; Franklin et al. 1980); however, Cheadle (1986) considered the sediments were not deposited in a rift but may have been subsequently preserved in a rift structure. The association of the Sibley Group with anorogenic granite and volcanic rocks in northern Lake Nipigon suggests that Sibley Group deposition may be controlled by sagging of the lithosphere during pre–Grenville extension and that the Sibley Group may be derived from rapid erosion of felsic volcanic rocks.

Mesoproterozoic Keweenawan Rifting

The Mesoproterozoic Midcontinent Rift occupies an arcuate belt over 2000 km long in the central part of North America (King and Zietz 1971; Chase and Gilmer 1973; Ocola and Meyer 1973; Halls 1978). Seismic reflection studies in Lake Superior have shown that the rift is a system of linked half grabens of alternating structural polarity (*see* Figure 16.26, Cannon et al. 1989). Volcanic rocks and sedimentary rocks that are up to 30 km thick (Cannon et al. 1989) fill the axial region of normal fault-bounded grabens. The normal rift-bounding faults are largely beneath Lake Superior and include the Douglas–Keweenaw and Isle Royale–Michipicoten Island faults. Accomodation zones bounding half grabens of structural alternating polarity are best exemplified by the Thiel fault which extends from the vicinity of the Slate Islands in Ontario to the Keweenaw Peninsula in Michigan. Had the rift developed further, structures such as the Thiel fault may have developed into transform faults. Late uplift of the central graben occurred when the original normal faults were reactivated as high-angle reverse faults (Cannon et al. 1989). Coincident with rifting, alkalic magmas were emplaced along major long-lived fracture zones such as the Kapuskasing Structural Zone and along the Trans–Superior Tectonic Zone, which extends north from the Thiel fault.

Stages in the rift development based on the seismic reflection data of Cannon et al. (1989) are indicated schematically in Figure 16.37. The earliest phase of rifting is characterized by the eruption of volcanic rocks over a wide area and is associated with broad crustal sag and subordinate normal faulting. Subsequent extension and crustal thinning results in the development of a distinct fault-bounded central graben and by the deposition of volcanic rocks largely in the central graben. In the late stages of rifting, thermal collapse results in the ending of volcanism and the onset of major sedimentation. Tectonic inversion of the rift is indicated by reverse movements on the early normal faults and by uplift of the central graben.

Based on the thickness of the volcanic-dominated section from seismic reflection profiles (Cannon et al. 1989), the volume of basaltic magmatism in the Keweenawan rift probably exceeds 10^6 km^3 and is one of the most voluminous basaltic sequences preserved on the earth. In Ontario, the thickest Keweenawan volcanic section is in eastern Lake Superior, where a supracrustal sequence in excess of 20 km thick south of Michipicoten Island is indicated by seismic reflection profiles (Cannon et al. 1989).

Petrological studies of intrusive rocks in the Nipigon area led Sutcliffe (1987) to suggest that relatively dense, olivine-phyric (picritic) magmas were emplaced early in the rift development along fractures that extended through a strongly coupled crust-mantle interface. The transition to more fractionated aphyric tholeiitic magmas may indicate the development of a magma holding chamber, possibly at the base of the crust. These relationships are depicted schematically in Figure 16.38. Subsequent studies of extrusive rocks in the Osler Group (Lightfoot et al., in press) and at Mamainse Point (Berg and Klewin 1988) documented a similar transition from olivine-phyric to aphyric and plagioclase-phyric basalts. Although the study of Lightfoot et al. (in press) does not preclude underplating at the base of the crust, geochemical constraints for contamination by a component with high Th-Yb and large-ion-lithophile elements and the apparent absence of high pressure fractionation trends suggest that upper crustal processes explain the compositional trends in aphyric or plagioclase-phyric magma types.

The role of an active versus passive mantle in the formation of the Midcontinent Rift is not resolved. Donaldson and Irving (1972), McWilliams and Dunlop (1978) and Gordon and Hempton (1986) argued that the rifting was caused by compression in the Grenville Province. Uranium-lead age determinations in the Grenville Province (Krogh and Wardle 1984; von Breeman et al. 1986) demonstrate that northwest-directed thrusting in the Grenville orogen at approximately 1.1 Ga was coincident with Keweenawan rifting.

The alternative model that rifting is related to an active mantle has been proposed by Burke and Dewey (1973). In this model, it is suggested that a plume-generated rift- rift-rift junction was positioned near the present north shore of Lake Superior. Based on the presence of 1.1 to 1.0 billion-year-old carbonatite-alkalic complexes and the prominent gravity high, Burke and Dewey (1973) suggested that the Kapuskasing Structural Zone may have been a failed rift arm. The Kapuskasing Structural Zone, however, has subsequently been interpreted as a west-dipping crustal-scale thrust (Percival and Card 1983). The Nipigon area may be a more reasonable candidate for a failed arm of the Midcontinent Rift as has been suggested by Card et al. (1972), Franklin et al. (1980) and Chandler et al. (1982). This area occurs at a major flexure in the rift, and is the site of voluminous mafic magmatism, normal faulting and graben formation.

Cannon et al. (1989) concluded that an anomalously hot asthenosphere existed beneath the rift and is the most likely cause of the voluminous magmatism. This hot asthenosphere enabled the ductile stretching of the Archean rocks beneath the rift.

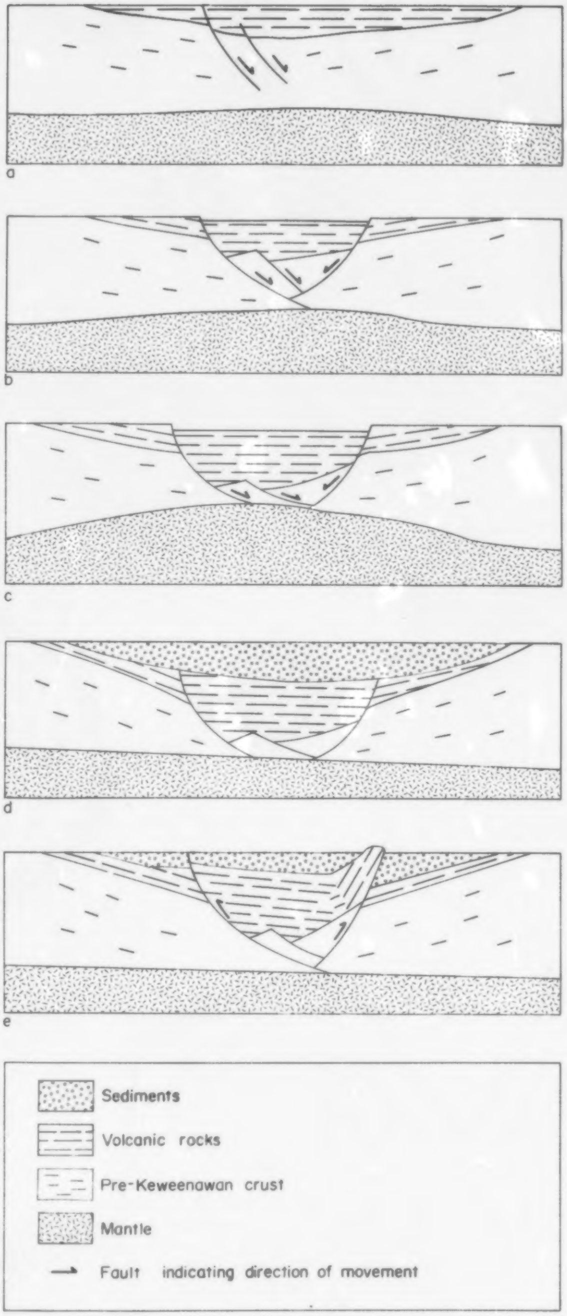

Figure 16.37. Cartoon depicting stages in the tectonic development of the Midcontinent Rift (*modified from* Cannon et al. 1989).

ECONOMIC GEOLOGY

The Keweenawan rift event is well known for basalt-hosted deposits of native copper on the Keweenaw Peninsula (White 1968) and for the shale-hosted White Pine deposit in Michigan (Ensign et al. 1968). In addition, iron ranges in the lower Animikie Group are an important deposit type in the Lake Superior region (Marsden et al. 1968). Mining of these types of deposits, however, has largely been confined to areas in the USA. In Ontario, with the exception of amethyst quarries near Thunder Bay, there are currently no producing mines in Proterozoic rocks of the Lake Superior region. Past production, however, includes copper from the Batchewana area (Grunsky 1991) and Michipicoten Island and silver from the Thunder Bay area (Sergiades 1968; Franklin et al. 1986). In addition, large tonnage, low-grade, copper-nickel resources with (PGE) values have been delineated in the Port Coldwell alkalic complex and Crystal Lake Gabbro. Uranium occurrences are also associated with unconformities between the Archean rocks and younger sequences.

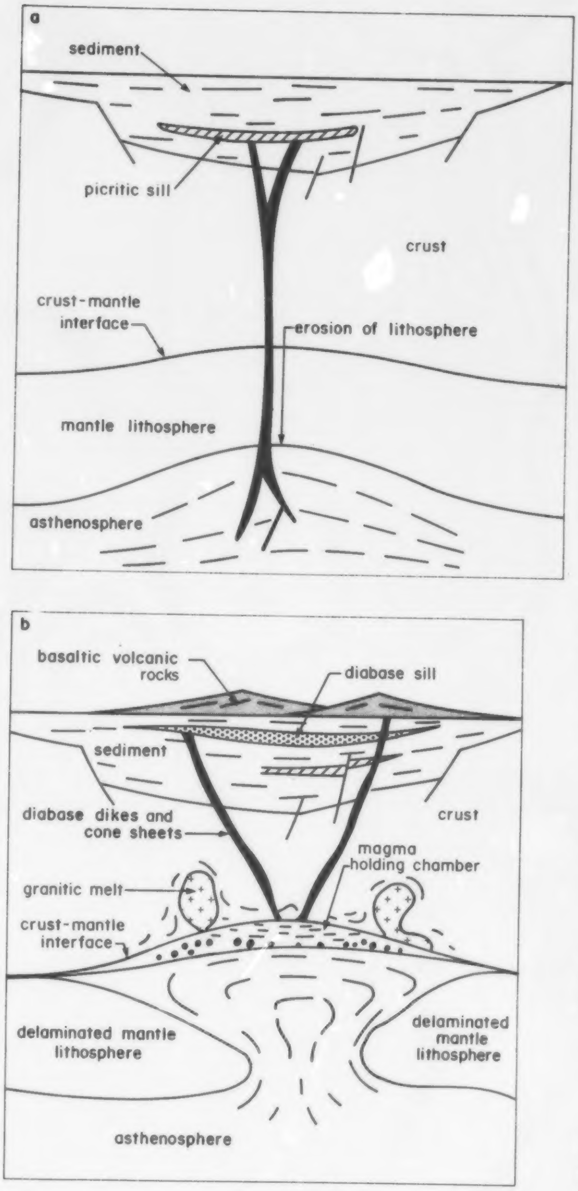

Figure 16.38. Schematic model for magmatism in the Nipigon area (*modified from* Sutcliffe 1987): **a**) emplacement of picritic magmas along fractures, early in rift development; and **b**) development of a magma holding chamber.

Figure 16.39 shows occurrences of mineralization of Keweenawan age in Ontario.

Magmatic Copper-Nickel Sulphide Mineralization

The Crystal Lake Gabbro hosts the Great Lakes Nickel deposit, a large tonnage, low-grade, copper-nickel deposit which also contains PGE values. The most recently published grades for this deposit are 0.334% Cu, 0.183% Ni, 0.02 ounce Pd per ton and 0.006 ounce Pt per ton (Postle et al. 1986). The copper-nickel is contained within pegmatitic gabbro in the lower part of the Crystal Lake Gabbro. Mineralization consists of disseminated chalcopyrite, pyrrhotite and pentlandite (Geul 1970). Chromite-bearing layers in the overlying layered middle zone (*see* Figure 16.25) also locally contain anomalous PGE values (Smith and Sutcliffe 1987). Whittaker (1986) suggested that the potential for economic chromite concentration is low due to the disseminated nature of the chromite in the layers.

The Crystal Lake Gabbro has lithological similarities to the Duluth Gabbro and both bodies are intruded into argillite and wacke of the Animikie Group. Assimilation of sulphur-rich sediments has been proposed to account for the origin of copper-nickel sulphide minerization at the base of the Duluth Complex (Weiblen and Morey 1980; Ripley 1981). Based on the presence of hornfelsic metasedimentary xenoliths in the basal part of the Crystal Lake Gabbro, a similar model may be appropriate for the Great Lakes Nickel deposit mineralization.

A copper and PGE deposit known as the Marathon deposit occurs in coarse-grained to pegmatitic gabbro on the eastern margin of the Port Coldwell alkalic complex. Drilling of the deposit has indicated 47 million tons of 0.02 ounce Pt per ton, 0.054 ounce Pd per ton, 0.42% Cu and traces of Ni, Co, Rh, Au and Ag (Dahl et al. 1986). The host gabbro consists of plagioclase, olivine, augite, orthopyroxene and biotite with apatite and oxides. The mineralization which consists of disseminated chalcopyrite-cubanite-pyrrhotite-pentlandite-pyrite is considered to have been influenced by assimilation of sulphur-rich Archean metavolcanic host rocks (Dahl et al. 1986) and by hydrous saline fluids (Watkinson et al. 1986).

A copper-gold-palladium occurrence known as the MacRae occurrence is also present along a gabbro-syenite contact within the interior of the Port Coldwell alkalic complex.

Copper Mineralization in the Keweenawan Supergroup

In the latter part of the nineteenth century, copper was produced from the Mamainse Point area (Thomson 1954) and from the Quebec mine at the west end of Michipicoten Island (Annells 1974). These early efforts were a result of the recognition of geological similarities to the Michigan copper district. More recently, copper has been mined from the Tribag and Coppercorp Ltd. mines in the Batchewana area, approximately 70 km northwest of Sault Ste. Marie (Giblin 1966).

At the Coppercorp Mine, chalcocite-rich mineralization occurs in fissure veins hosted by Keweenawan basalt of the Mamainse Point Formation. The veins are composed of quartz-calcite gangue and in addition to chalcocite, contain lesser bornite, chalcopyrite, and native copper (Giblin 1966). The Coppercorp Ltd. Mine produced 1 294 297 tons of ore with 2.1% Cu (Grunsky 1991). The Tribag copper-silver mine, located approximately 15 km northeast of the Coppercorp Ltd. Mine, occurs in one of several breccia

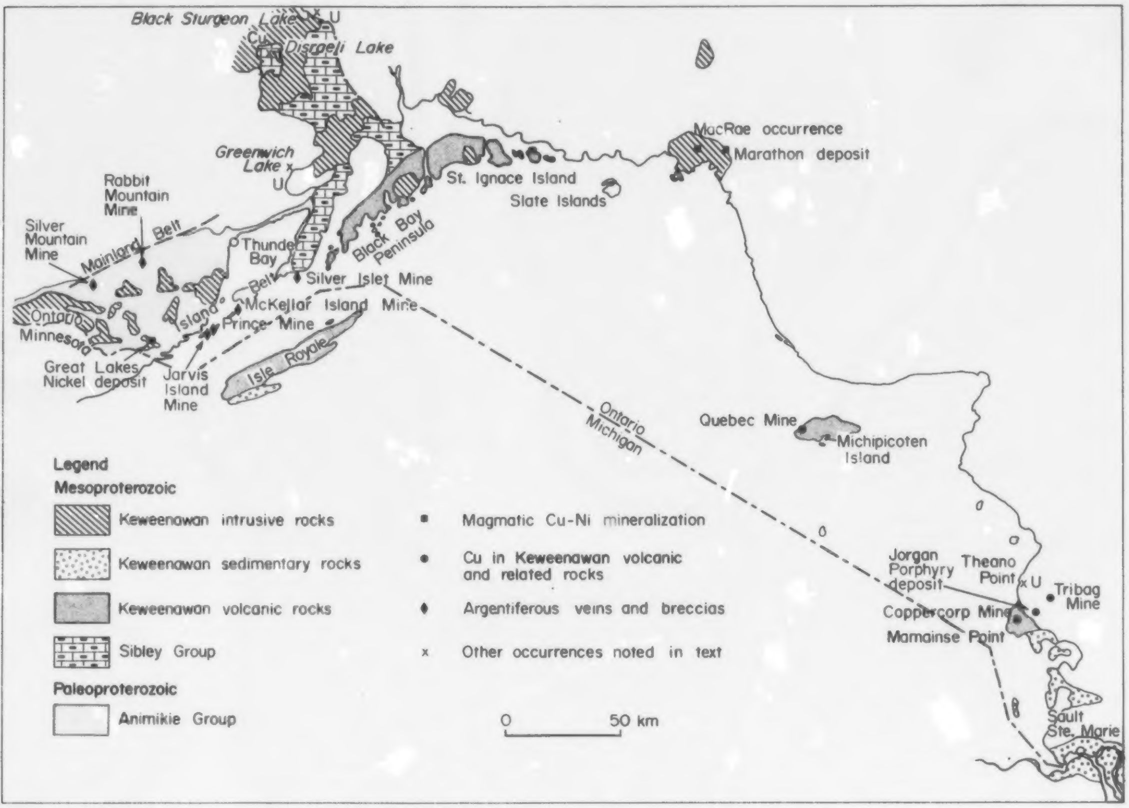
Black Sturgeon Lake
U
Cu
Disraeli Lake
Greenwich Lake
U
Rabbit Mountain Mine
Silver Mountain Mine
Mainland Belt
Thunder Bay
Belt
Island
Ontario
Minnesota
Great Lakes Nickel deposit
Jarvis Island Mine
McKellar Island Mine
Prince Mine
Silver Islet Mine
Isle Royale
Black Bay Peninsula
St. Ignace Island
Slate Islands
MacRae occurrence
Marathon deposit
Ontario
Michigan
Quebec Mine
Michipicoten Island
Jorgan Porphyry deposit
Theano Point
U
Tribag Mine
Coppercorp Mine
Mamainse Point
Sault Ste. Marie
Legend
Mesoproterozoic
Keweenawan intrusive rocks
Keweenawan sedimentary rocks
Keweenawan volcanic rocks
Sibley Group
Paleoproterozoic
Animikie Group
Magmatic Cu-Ni mineralization
Cu in Keweenawan volcanic and related rocks
Argentiferous veins and breccias
Other occurrences noted in text
0
50 km

Figure 16.39. Locations of major mineral occurrences of probable Keweenawan age in Ontario.

pipes of probable Keweenawan age which are intrusive into Archean rocks. The oval-shaped Tribag breccia pipe has dimensions of approximately 100 by 400 m and consists of angular fragments of granitoids, mafic metavolcanic rocks, felsite and diabase in a matrix of vuggy quartz-carbonate (Blecha 1965; Armbrust 1969). Mineralization consists of chalcopyrite, pyrite, molybdenite, sphalerite and galena and is associated with chlorite, sericite and clay mineral alteration (Blecha 1965). The alteration is dated at 1050 Ma by the K-Ar method (Armbrust 1969). The Breton Pipe, which represents the main Tribag ore zone, produced 37 257 993 pounds of Cu and 246 054 ounces of Ag (Grunsky 1991). The west breccia zone of the Tribag Mine contains 28 000 tons of 0.87% WO_3 and 40 000 tons of 2.0% Cu, although no ore has been removed (Grunsky 1991).

The Jogran Porphyry deposit, located in Ryan Township at the north end of Mamainse Lake, contains disseminated pyrite, chalcopyrite, molybdenite and chalcocite mineralization with silver values. Richards and Spooner (1986) considered the origin of the Jogran and Coppercorp Ltd. deposits to be a result of magmatic fluids related to porphyry copper intrusions of Keweenawan age.

Native copper mineralization in amygdaloidal basalt and copper-chalcocite mineralization in carbonate veins that cut basalt are also present in the St. Ignace Island area (Giguere 1975), on Black Bay Peninsula (McIlwaine and Wallace 1976), on Michipicoten Island (Annells 1974) and on the Slate Islands (Sage 1983).

Copper occurrences at Disraeli Lake, south of Lake Nipigon, are associated with Nipigon diabase sills and the underlying stromatolitic carbonates of the Sibley Group. Coates (1972) considered these occurrences to be related to diabase intrusion and to later supergene enrichment processes.

Silver Veins

Argentiferous calcite-barite veins and calcite-quartz and locally amethyst-filled breccia zones occur in the vicinity of the north shore of Lake Superior near Thunder Bay. In the latter part of the nineteenth century, mining operations for silver took place at Silver Islet, Silver Mountain, Rabbit Mountain and the Prince Mine. The exploitation of these deposits during the period 1868 to 1890 tranformed the Thunder Bay area from a sparsely settled wilderness, dependant solely on the fur trade, to an area with a diverse economic base. In addition, barite was produced at McKellar Island and Jarvis Island. At Silver Islet, silver mineralization occurs as native silver, as well as argentite, niccolite, galena, sphalerite, marcasite, cobaltite, smaltite, domeykite, chalcopyrite and tetrahedrite in a gangue of carbonate and quartz (Sergiades 1968). Mineralization is associated with a northwest-striking cross fault in an olivine diabase dike (Sergiades 1968).

Franklin et al. (1986) subdivided the silver deposits into 3 groups. The Mainland veins mainly occur in Rove Formation shales below the contact with Logan diabase sills and are associated with a zone of normal faults near the western margin of Proterozoic rocks. The Island group of veins occur in close proximity to northeast-trending gabbro and diabase dikes and are hosted by both the Rove Formation shales and dikes. The veins mainly occur in fractures perpendicular to the dike hosts. A third group of veins occur near Port Coldwell and in a shear zone that cuts Archean metasedimentary rocks and a Proterozoic diabase dike. Franklin et al. (1986) considered that both the Island and Mainland groups were deposited in structures formed dominantly by normal faulting during the late stages of rifting with heat supplied by mafic magmatism and metals derived from both shales and mafic rocks.

Uranium Occurrences

Uranium mineralization has been documented by Franklin (1978) and is associated with northwest-trending fracture zones related to the Black Sturgeon graben structure. Mineralization of this type at Greenwich Lake and south of Black Sturgeon Lake occurs in fractures in Archean rocks near the Archean–Proterozoic unconformity. Ruzicka and LeCheminant (1984) dated the uranium minerals in fractures associated with the Black Sturgeon fault at 1094 Ma, which suggests that the mineralization is Keweenawan in age. Uranium mineralization is also present along the east shore of Lake Superior at Theano Point north of Mamainse Point (Giblin and Leahy 1965).

CONCLUSIONS

The Lake Superior area was the site of cover sequences and intrusive activity from 1.86 Ga through to *circa* 1 Ga. This activity included: development of Paleoproterozoic Animikie Group foredeep succession related to the 1.86 Ga Penokean Orogeny; Mesoproterozoic anorogenic magmatism and Sibley Group deposition related to sagging during the pre–Grenville extension; and the voluminous volcanic and sedimentary rocks of the Keweenawan Supergroup, confined to the 1.1 billion-year-old Midcontinent Rift. Mineral deposits of the area reflect the variety of supracrustal depositional environments and the various types of igneous episodes over the lengthy, active history of the region.

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Chapter 17

Proterozoic Mafic Dike Swarms in the Superior Province of Ontario

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Abstract

More than 12 Precambrian mafic dike swarms, which range in age from 2.5 to 1.1 Ga, occur in the Superior Province of the Canadian Shield in Ontario. Major dike swarms described in this chapter are the Matachewan and Hearst, Kenora–Fort Frances, Preissac, Marathon, and Molson swarms of Paleoproterozoic age; and the Mackenzie, Sudbury, Abitibi and Keweenawan (Pukaskwa, Pigeon River and Nipigon) swarms of Mesoproterozoic age. Most of these swarms are considered to have been formed in a tensional tectonic setting associated with intraplate rifting or plate margin activity.

The majority of the mafic dike swarms in the Superior Province are tholeiites. The Mesoproterozoic dikes are enriched in incompatible elements relative to the Paleoproterozoic dikes. Trace element abundances in Mesoproterozoic swarms show similarity with the Phanerozoic continental flood basalts and oceanic-island basalts, whereas the Paleoproterozoic swarms have some resemblance to calc-alkalic basalts.

Coeval mafic intrusions including the Nipissing, Nipigon and Sutton Inlier diabases are also briefly discussed in this chapter because their tectonic setting is closely related to correlative mafic dike swarms.

INTRODUCTION

The Superior Province of Ontario hosts a large number of Paleoproterozoic to Mesoproterozoic mafic dikes (Figure 17.1) which have been separated into a number of swarms based on petrographic, geochemical, paleomagnetic and absolute age criteria (Fahrig et al. 1965; Gates and Hurley 1973; Ernst 1981; Card 1981; Condie et al. 1987; Halls and Shaw 1987). The objective of this chapter is to summarize information on the distribution, age and petrochemistry of the dike swarms, and also to review concepts regarding the tectonic settings of the dike swarms. The dike swarms reviewed here are restricted to those formed through injection of basaltic magma between 2.5 and 1.1 Ga (Table 17.1). Ultramafic, felsic and lamprophyre dikes are excluded. Dike swarms included in this review are the Matachewan and Hearst, Kenora–Fort Frances, Preissac, Marathon, Molson, Sudbury, Mackenzie, Abitibi and Keweenawan dike swarms. Their compositions, ages, tectonic environments and other physical aspects are summarized in Table 17.1. Coeval mafic intrusions such as Nipissing, Nipigon and Sutton Inlier diabases are also briefly discussed in this chapter because their tectonic setting is closely related to correlative mafic dike swarms. Alkalic and carbonatitic intrusions, although shown in Figure 17.1, are not discussed in the text. Details regarding these intrusive rocks can be found elsewhere in the volume (Sage, this volume).

The term "swarm", in its formal sense, defines any single concentration of dikes which were emplaced during the same igneous episode (Speight et al. 1982). Detailed paleomagnetic, petrochemical and geochronological studies, which are necessary to distinguish the various dike swarms, have not been completed on all swarms in Ontario; therefore, in some cases, dikes in a single concentration may be of several ages.

The geological record shows that the margin and interior of the Archean Superior Province was reactivated in a series of major magmatic and tectonic events during the Paleoproterozoic and Mesoproterozoic. Evidence for these events includes: 1) reactivation of major Neoarchean transcurrent shear zones and faults (2.7 to 1.9 Ga, Peterman and Day 1989; Osmani et al. 1989; Kamineni et al. 1990); 2) brittle faulting (2.3 to 2.2 Ga, Kamineni et al. 1990); 3) formation of sedimentary basins such as the one occupied by the Huronian Supergroup (*see* Bennett et al., this volume); 4) emplacement of mafic dike swarms (2.45 to 1.1 Ga, *see* Table 17.1); and 5) intrusion of carbonatite and alkalic magmas (*see* Sage, this volume). Some of these events, particularly the development of large basins and regional dike swarms, have been linked to the rifting of the Archean craton and to the suturing of Proterozoic terranes onto the margins of the craton (Baragar and Scoates 1981; Fahrig 1987; Hoffman 1988; Gibb 1983).

An alternative tectonic model of dike swarm emplacement is one related to convergent tectonism. In this model, dikes are intruded normal to the collision zone and parallel to the maximum compressive stress (Feraud et al. 1987). In this review, the three-stage plate-tectonic cycle of Fahrig (1987), is used to explain the development of most mafic dike swarms in the Superior Province of Ontario. The proposed model (Figure 17.2) of Fahrig (1987) involves the following sequence:

1. the formation of a zone of tension with dike intrusion, accompanied by rifting and volcanism
2. a stage of spreading which results in the development of a dike group parallel to the continental margin (passive

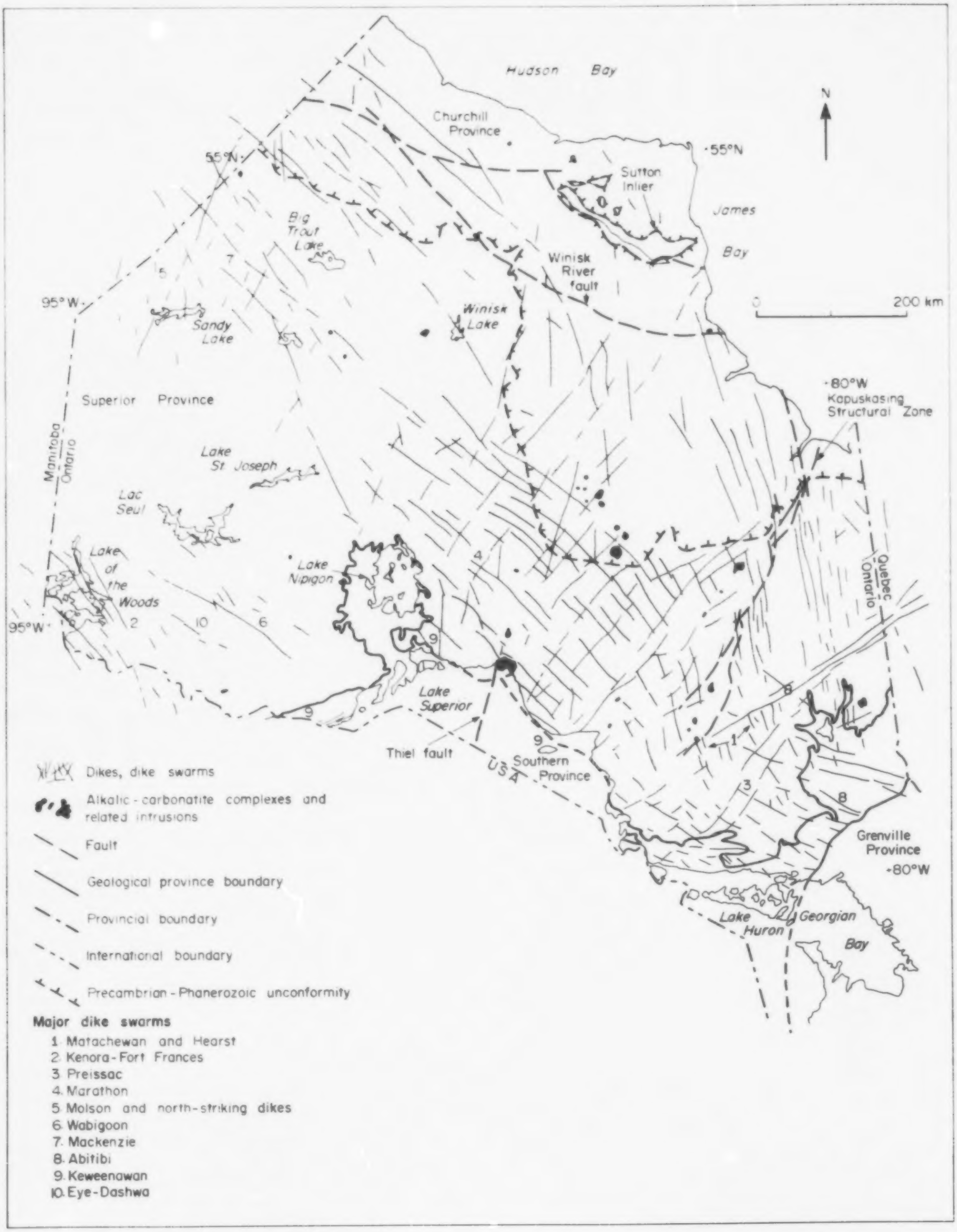
Hudson Bay
Churchill Province
55°N
55°N
Sutton Inlier
James Bay
N
Big Trout Lake
Winisk River fault
95°W
Sandy Lake
Winisk Lake
0
200 km
80°W
Kapuskasing Structural Zone
Superior Province
Manitoba
Ontario
Lake St. Joseph
Lac Seul
Lake of the Woods
Lake Nipigon
Quebec
Ontario
95°W
Lake Superior
Thiel fault
USA
Southern Province
Grenville Province
80°W
Lake Huron
Georgian Bay
1
2
3
4
5
6
7
8
9
10
Dikes, dike swarms
Alkalic - carbonatite complexes and related intrusions
Fault
Geological province boundary
Provincial boundary
International boundary
Precambrian - Phanerozoic unconformity
Major dike swarms
1. Matachewan and Hearst
2. Kenora - Fort Frances
3. Preissac
4. Marathon
5. Molson and north-striking dikes
6. Wabigoon
7. Mackenzie
8. Abitibi
9. Keweenawan
10. Eye - Dashwa

Figure 17.1. Map showing the general distribution of Proterozoic (2.5 to 1.1 Ga) dike swarms, alkalic-carbonatitic complexes (*see* Sage, this volume) and other elements in the Superior Province of Ontario.

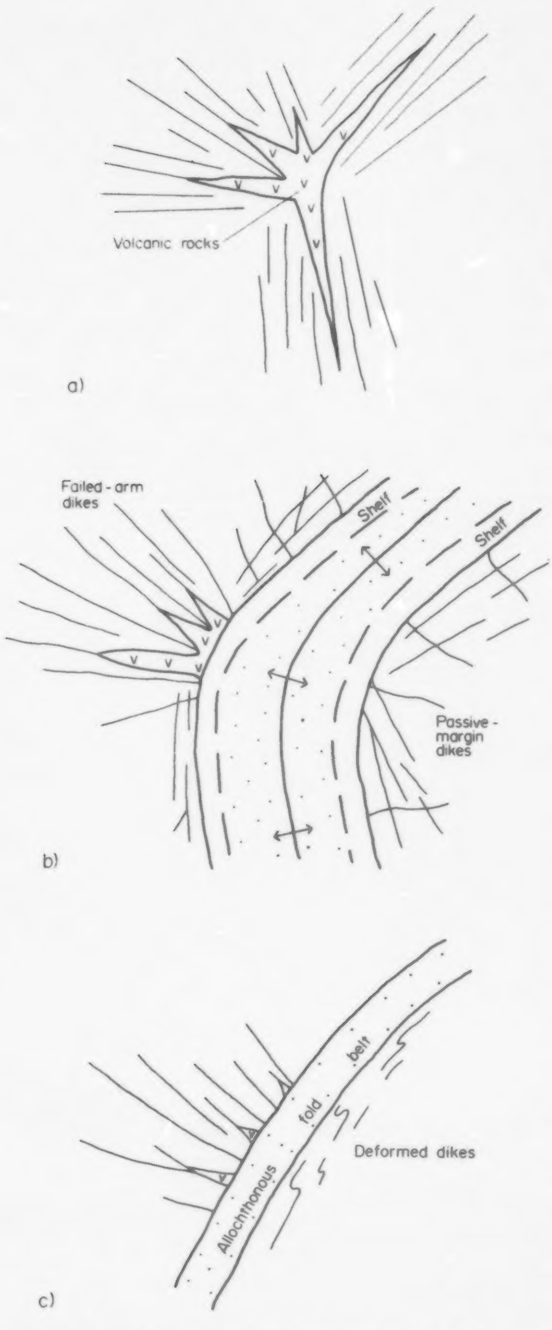

Figure 17.2. Proposed tectonic stages in the development of mafic dike swarms in the Superior Province (*modified from* Fahrig 1987). **a)** Dike intrusion, rifting and volcanism. **b)** Ocean opening. **c)** Ocean closing.

margin) and a dike group occupying a failed arm in the interior of the continents

3. In the event of ocean closure, the passive-margin rifts and dikes will be consumed, leaving only failed-arm dikes, volcanic rocks and rifts that extend into the continent.

For this review, and for the Bedrock Geology of Ontario maps (*see* Maps 2541 to 2545, map case), the locations of dikes were compiled using existing compilation maps, reports, Landsat imagery, and standard and high resolution (lower flight altitudes and closer flight line spacings) federal and provincial aeromagnetic maps. The distribution of Proterozoic dike swarms in the Superior and Southern provinces is shown on Figure 17.1.

The criteria used to identify mafic dikes on aeromagnetic maps are: 1) linear, narrow positive or negative anomalies; and 2) the crosscutting aspect of the anomalies with respect to local host rock trends. Anomalies parallel to host-rock foliation trends are interpreted with less certainty than those which cut them at some angle.

Linear anomalies are quite distinct in areas occupied by relatively low magnetic susceptibility granite, gneiss and metasedimentary rocks, though they may be obscure in areas occupied by strongly ferromagnetic country rocks such as iron formations, mafic to ultramafic extrusive and intrusive rocks, and some magnetite-bearing granitic rocks. Dike anomalies in areas covered by Phanerozoic rocks can be recognized in many localities and are also included in Figure 17.1.

MAFIC DIKE SWARMS

Matachewan and Hearst Dike Swarms

The Matachewan and Hearst dike swarms, in the central and southern Superior Province (*see* Figures 17.1, 17.3a and 17.3b), trend north to northwesterly and have a U-Pb age of 2454±2 Ma (Heaman 1988). The dike swarm, covering an area of 500 by 700 km (Halls 1988), is the second largest in the Canadian Shield.

The dikes of the swarm are best known southeast of the Kapuskasing Structural Zone. Dikes of this swarm located northwest of the Kapuskasing Structural Zone have been largely correlated with those southeast of the Kapuskasing Structural Zone on the basis of strike, petrography, aeromagnetic data and paleomagnetic characteristics. The dikes are generally around 10 m across, with only a few exceeding 250 m, and have vertical to subvertical dips (Condie et al. 1987). They are mainly quartz diabase and are commonly characterized by blocky phenocrysts of saussuritized plagioclase up to 20 cm in length (Phinney and Morrison 1988). The groundmass is composed of plagioclase (50 to 60%), augite (30%), quartz (5 to 10%) and opaques (5%)(Ernst 1981). Other accessory minerals include apatite, pyrite and pyrrhotite (Ernst 1981).

The dikes of the Matachewan and Hearst dike swarms are not known to occur in the granulite-facies rocks of the Kapuskasing Structural Zone, though they are commonly identified on either side of the zone (Ernst 1981). This apparent lack of dikes in Kapuskasing Structural Zone can be explained by the suggestion that they were emplaced prior to major strike-slip and thrust faulting, that is, uplift of mid-crustal rocks (1.9 Ga)(Percival and Card 1983) which were below the maximum depth of dike intrusion. However,

Table 17.1. Paleoproterozoic and Mesoproterozoic (2.5 to 1.1 Ga) dikes and dike swarms within the Superior Province of Ontario and adjacent areas.

Dike Swarm	General Strike	General Distribution	Geological Data	Petrology	Tectonic Environment	Age; References
Matachewan and Hearst	N-NW	Northeast, central and northwestern Ontario, 500 by 700 km	Intrudes Archean; cut by Marathon, Preissac, Sudbury and Abitibi dikes; intermixed with similar-striking Mackenzie dikes in northwestern Ontario; vertical to subvertical dip	[1]Quartz diabase typically porphyritic (up to 20 cm plagioclase phenocrysts) [2]Quartz tholeiite	Failed arm/passive margin	[3]2454 ± 2 Ma; Heaman (1988)
Kenora–Fort Frances	NW	Mesabi Iron Range (Minnesota, USA) to Kenora (Ontario, Canada), 300 by 300 km	Intrudes Archean; deformed by Penokean Orogeny; variable dips	[1]Quartz diabase; hypidiomorphic granular texture [2]Iron-enriched quartz tholeiite	Predominantly failed arm/foredeep basin	[4]2120 ± 67 Ma; Beck and Murthy (1982) [5]2240 Ma; Hanson and Malhotra (1971)
Preissac	NE	Northeast of Lake Superior	Intrudes Archean and early Proterozoic; cut by E-NE-striking Abitibi dikes	[1]Quartz diabase [2]Quartz tholeiite	Failed arm/foredeep basin	[4]2150 Ma; Gates and Hurley (1973)
Marathon	NE	North of Lake Superior	Intrudes Archean; intermixed with Nipigon and Preissac dikes	[1]Quartz and olivine diabase	Failed arm	[5]1964 Ma; Wanless et al. (1966) [6]2170 Ma; Fahrig and West (1986)
Wabigoon[7]	NW	Wabigoon Lake and Lac des Mille Lacs areas				[5?]1900 Ma; Fahrig and West (1986)
Molson	N-NE	Northwestern Superior Province	Intrudes high-grade gneisses of sequences of Sachigo Subprovince; subvertical dip	[2]Tholeiitic-komatiitic basalt	Failed arm/thrusting	[3]1883 ± 2 Ma; Heaman et al. (1986)
Sudbury	W-NW	Sudbury area	Intrudes Archean and early Proterozoic; vertical dip	[1]Olivine diabase [2]Alkali olivine basalt	Failed arm/oceanic spreading/Grenville collision	[3]1238.5 ± 4 Ma; Krogh et al. (1987)
Mackenzie	NW	Arctic coast near Coronation Gulf to northwestern Ontario, 1800 by 2400 km	Intrudes Archean and Proterozoic; subvertical dip	[1]Quartz (+ olivine) diabase [2]Quartz basalt	Failed arm/passive margin	[3]1267 ± 2 Ma; LeCheminant and Heaman (1989)
Abitibi	E-NE	Northeast of Lake Superior	Intrudes Archean and early Proterozoic; subvertical dip	[1]Olivine diabase [2]Alkali olivine basalt	Failed arm	[3]1140.6 ± 2 Ma; Krogh et al. (1987)
Pukaskwa	NW	Northeast shore of Lake Superior	Roughly parallel to the trend of Midcontinent Rift dip of 50° to 80° NE	[1]Olivine diabase	Passive margin	[8]1100 Ma

Table 17.1. Paleoproterozoic and Mesoproterozoic (2.5 to 1.1 Ga) dikes and dike swarms within the Superior Province of Ontario and adjacent areas.

Dike Swarm	General Strike	General Distribution	Geological Data	Petrology	Tectonic Environment	Age; References
Pigeon River	NE	Northwest shore of Lake Superior	Roughly parallel to the trend of Midcontinent Rift; dip steeply to SE; intrudes early to middle Proterozoic and Archean	[2]Olivine and quartz (minor) tholeiite	Passive margin	[8]1100 Ma
Nipigon	N	Along north-central shore of Lake Superior	Radiate from centre of the Lake Superior Basin; intrudes Proterozoic (Keweenawan igneous rocks and Sibley sedimentary rocks) and Archean; subvertical dip	[1]Olivine and quartz diabase [2]Olivine and quartz tholeiite	Failed arm	[8]1100 Ma
Other dikes[7]						
Pickle Crow	N-NW	North of Lake Nipigon to North Caribou–Muskrat Dam lakes areas	Intrudes Archean Wabigoon, English River, Uchi, Berens River and Sachigo subprovinces; roughly 460 km long and up to 90 m wide; dips steeply (75° to 85°)	[1]Quartz diabase		[6]>2000 Ma; C. Hale (personal communication)
Ear Falls	NW	Ear Falls–Red Lake areas	Intrudes Archean; generally narrow (<1 m), but few are several metres wide; few dikes are composite indicating multiple injection of magma			[6]1700 to 1900 Ma; Osmani (1982), Symons et al. (1983)
Eye–Dashwa	NW	Eye–Dashwa lakes, near Atikokan	Intrudes Eye–Dashwa Lake pluton; dikes probably belong to Keweenawan event		Failed arm	[5]1132 ± 27 Ma, 1143 ± 27 Ma; Hunt and Roddick (1987)

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[1] Petrography

[2] Chemistry

[3] U-Pb

[4] Rb-Sr

[5] K-Ar

[6] Paleomagnetic

[7] Dikes are not discussed in the text due to insufficient information available on them

[8] Geological interpretation

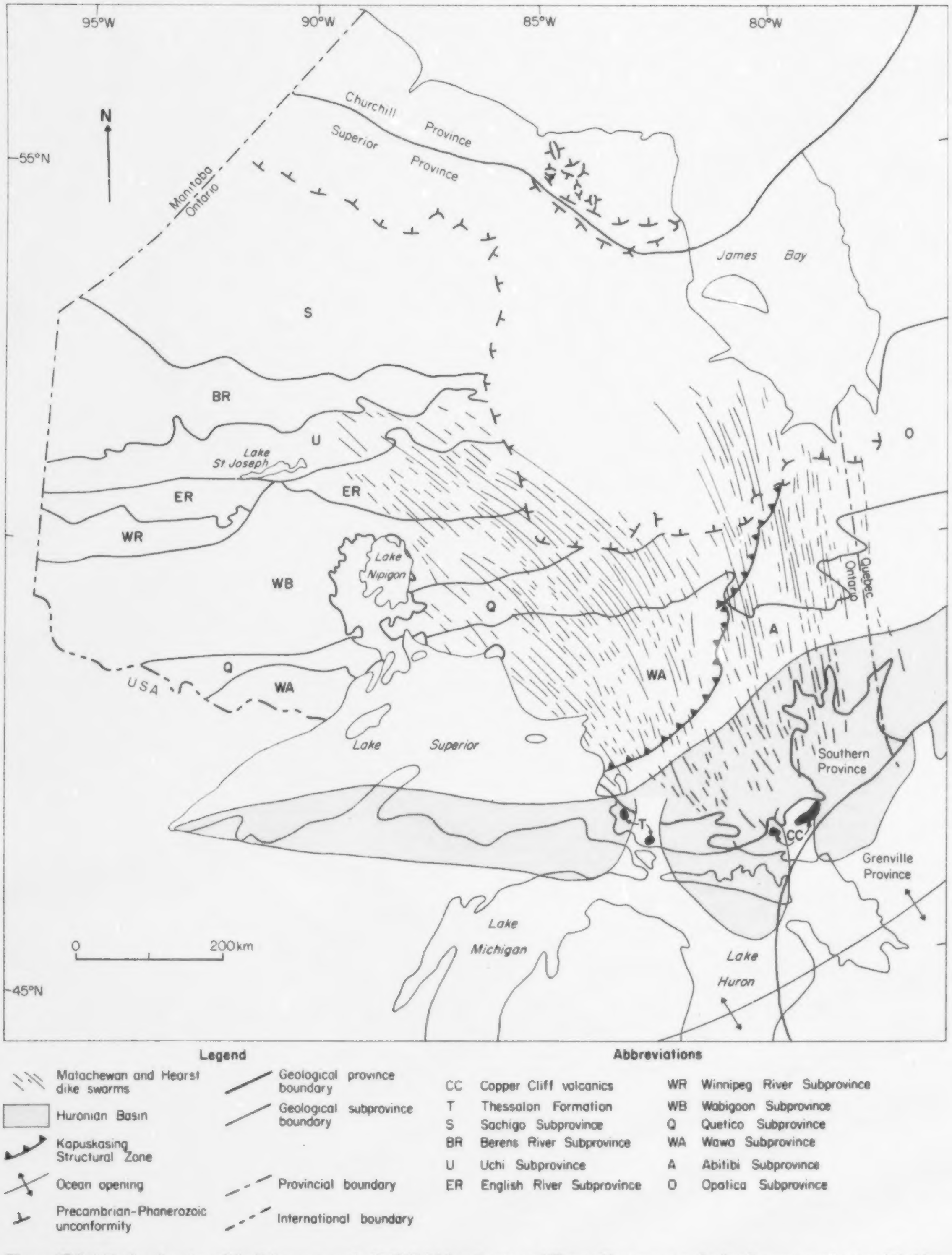

95°W
90°W
85°W
80°W
55°N
45°N
N
Manitoba
Ontario
Churchill Province
Superior Province
James Bay
S
BR
U
Lake St Joseph
ER
WR
WB
Lake Nipigon
Q
WA
A
O
Quebec
Ontario
USA
Lake Superior
Southern Province
T
CC
Grenville Province
Lake Michigan
Lake Huron
0
200km
Legend
Matachewan and Hearst dike swarms
Huronian Basin
Kapuskasing Structural Zone
Ocean opening
Precambrian-Phanerozoic unconformity
Geological province boundary
Geological subprovince boundary
Provincial boundary
International boundary
Abbreviations
CC Copper Cliff volcanics
T Thessalon Formation
S Sachigo Subprovince
BR Berens River Subprovince
U Uchi Subprovince
ER English River Subprovince
WR Winnipeg River Subprovince
WB Wabigoon Subprovince
Q Quetico Subprovince
WA Wawa Subprovince
A Abitibi Subprovince
O Opatica Subprovince

Figure 17.3.a) Regional setting of the Paleoproterozoic (2.45 Ga) Matachewan and Hearst dike swarms and related ocean opening (*modified from* Fahrig 1987). Extension of the Huronian Basin boundary as shown, is hypothetical and thought to have developed on the rifted margin of the Superior Province between 2450 and 2220 Ma (Kamineni et al. 1990). Also note that the major phase of thrusting in Kapuskasing Structural Zone is interpreted to predate the emplacement of the Matachewan and Hearst swarms.

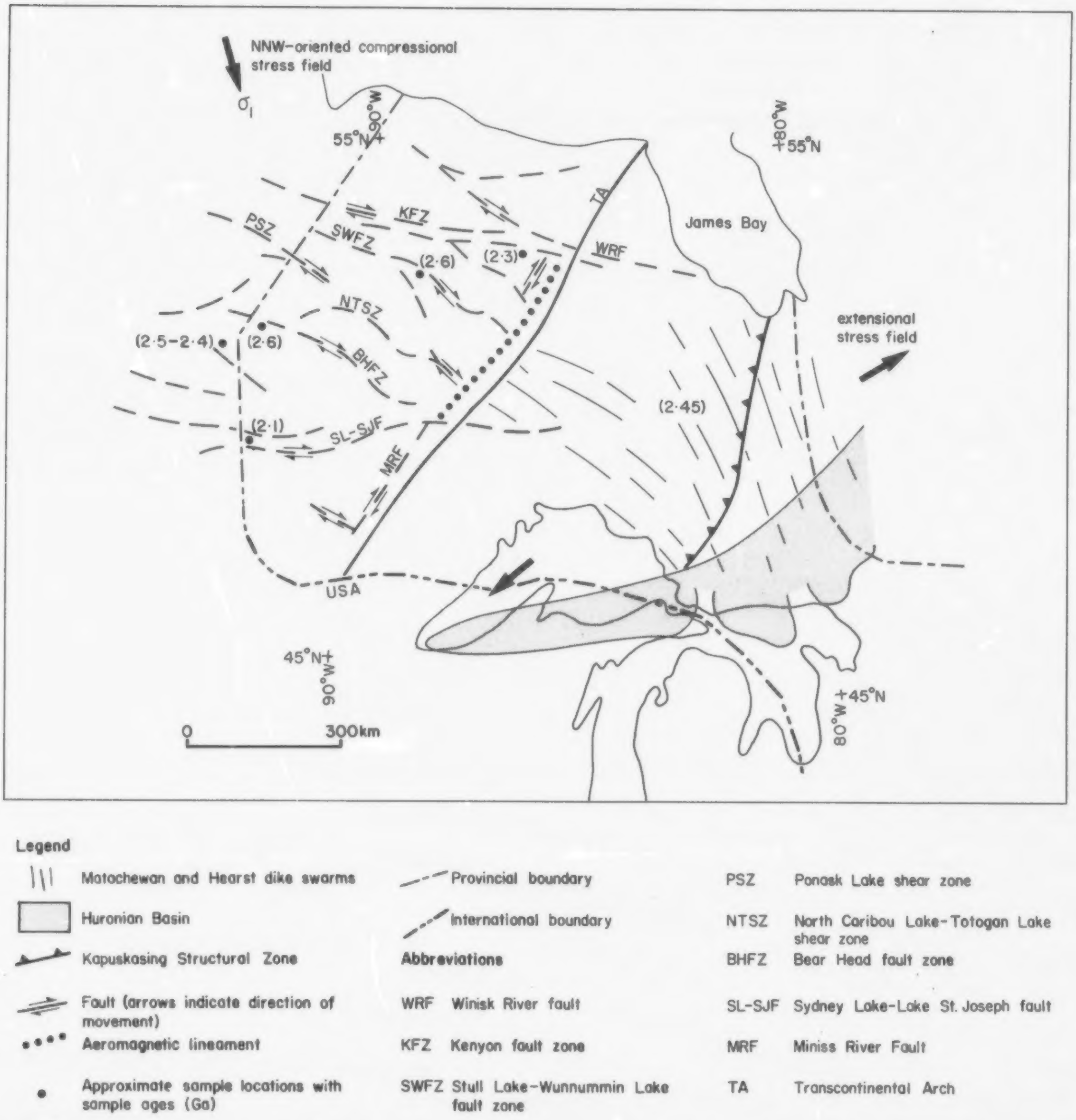

NNW-oriented compressional stress field
σ1
55°N
90°W
KFZ
PSZ
SWFZ
(2·6)
(2·3)
TA
WRF
James Bay
80°W
55°N
NTSZ
(2·5–2·4)
(2·6)
BHFZ
extensional stress field
(2·45)
(2·1)
SL-SJF
MRF
USA
45°N
90°W
0
300km
45°N
80°W
Legend
Matachewan and Hearst dike swarms
Huronian Basin
Kapuskasing Structural Zone
Fault (arrows indicate direction of movement)
Aeromagnetic lineament
Approximate sample locations with sample ages (Ga)
Provincial boundary
International boundary
Abbreviations
WRF Winisk River fault
KFZ Kenyon fault zone
SWFZ Stull Lake–Wunnummin Lake fault zone
PSZ Ponask Lake shear zone
NTSZ North Caribou Lake–Totogan Lake shear zone
BHFZ Bear Head fault zone
SL-SJF Sydney Lake–Lake St. Joseph fault
MRF Miniss River Fault
TA Transcontinental Arch

Figure 17.3.b) The proposed synchronous stress patterns directed on northwestern and south-central margins of the Superior Province between 2.68 and 2.1 Ga. Ocean spreading and associated east-west extension, which caused the emplacement of Matachewan and Hearst dike swarm in the south-central Superior Province, is possibly counteracted in the northwest by a contemporaneous north- to northwest-oriented compressional stress regime (σ_1). Numbers in brackets represent sample ages (Ga) determined by K-Ar and Rb-Sr techniques.

recent mapping (Leclair and Poirier 1989) and data from aeromagnetic surveys interpreted by this author, suggest the Matachewan and Hearst dikes do occur in the granulite-facies rocks within the Kapuskasing Structural Zone.

Three distinct dike sets within the Matachewan and Hearst dike swarms, the eastern, central and western dike sets, have been recognized south-southeast of the Kapuskasing Structural Zone. The eastern and central dike sets are relatively straight and, although they are interrupted by the Kapuskasing Structural Zone, they show no evidence of major lateral displacement since the time of intrusion. Although most dikes within the western set are bent around the southwestern end of the Kapuskasing Structural Zone, they show no significant lateral disruption of their aeromagnetic anomalies. It is also interesting to note that several north- and northwest-striking dikes, which obliquely cut the trend of the granulites and the faults, are straight and continuous across the Kapuskasing Structural Zone (*see* Figure 17.1). Hence, this crosscutting relationship between the dikes and the tectonic fabric of the Kapuskasing

Structural Zone suggests that some thrusting may have preceded dike emplacement.

The Matachewan and Hearst dikes extend north as far as the Uchi–Berens River subprovince boundary (*see* Figure 17.3a). Four possible scenarios which might explain the absence of Matachewan and Hearst dikes in this part of northwestern Ontario are as follows:

1. The northwestern limit of the Matachewan–Hearst magma chamber was simply coincident with the Uchi–Berens River subprovince boundary; thus the dike distribution does not require a specific geological model.
2. The northwestern limit of the Matachewan and Hearst dike swarms is related to known fault zones in the Miminiska Lake and Lake St. Joseph areas (*see* Figure 17.3b).
3. The termination of this dike swarm in northwestern Ontario may have been the result of synchronous variations in the stress patterns throughout the Superior craton during the emplacement of the Matachewan and Hearst dike swarms (*see* Figure 17.3b). Simultaneous with the east-west extensional tectonism (associated with the Matachewan and Hearst dike swarm emplacement in the south-central Superior Province) a regional north- to northwest-oriented compressional stress field related to Paleoproterozoic faulting (Peterman and Day 1989; Kamineni et al. 1990)(*see* Figure 17.3b) was operating in the northwestern Superior Province. This compressional event possibly prevented the propagation of the Matachewan and Hearst dike swarms in that area.
4. The Berens River Subprovince described by Card and Ciesielski (1986) was subjected to higher pressure metamorphism than the Uchi Subprovince (*see* Thurston et al., this volume), suggesting the Berens River Subprovince may represent mid–crustal depths. Therefore, the lack of Matachewan and Hearst dikes may reflect a limited depth extent for this swarm in the Berens River Subprovince.

The Matachewan and Hearst dike swarms are bounded to the south by rocks of the Huronian Supergroup. The dikes are the same age as the Copper Cliff felsic volcanic rocks (2450^{+25}_{-10}Ma, Krogh et al. 1984) and slightly younger than layered gabbro intrusions such as the East Bull Lake Gabbro-Anorthosite Intrusion (2480^{+10}_{-5}Ma, Krogh et al. 1984).

The southern convergence of the Matachewan and Hearst dike swarms is consistent with a failed-arm environment with a spreading point to the south (*see* Figure 17.3a and 17.3b). Fahrig (1987) relates emplacement of the Matachewan and Hearst dike swarms to the opening of an ocean prior to the development of the Huronian Supergroup. The Copper Cliff Formation (2450^{+25}_{-10}Ma) and correlative Thessalon Formation(Krogh et al. 1984), and layered intrusions are interpreted to represent a passive-margin, rift-associated igneous suite related to the Matachewan and Hearst dikes.

Kenora–Fort Frances Dike Swarm

The northwest-trending Kenora–Fort Frances dike swarm has an Rb-Sr age of 2120±67 Ma (Beck and Murthy 1982; Halls 1986; Southwick and Halls 1987) and extends from the Mesabi Iron Range in Minnesota, USA, to Kenora, Ontario (Figure 17.4). The dike swarm contains hundreds of dikes up to 100 km long and 120 m wide, covering an area of 90 000 km^2 (Southwick and Halls 1987).

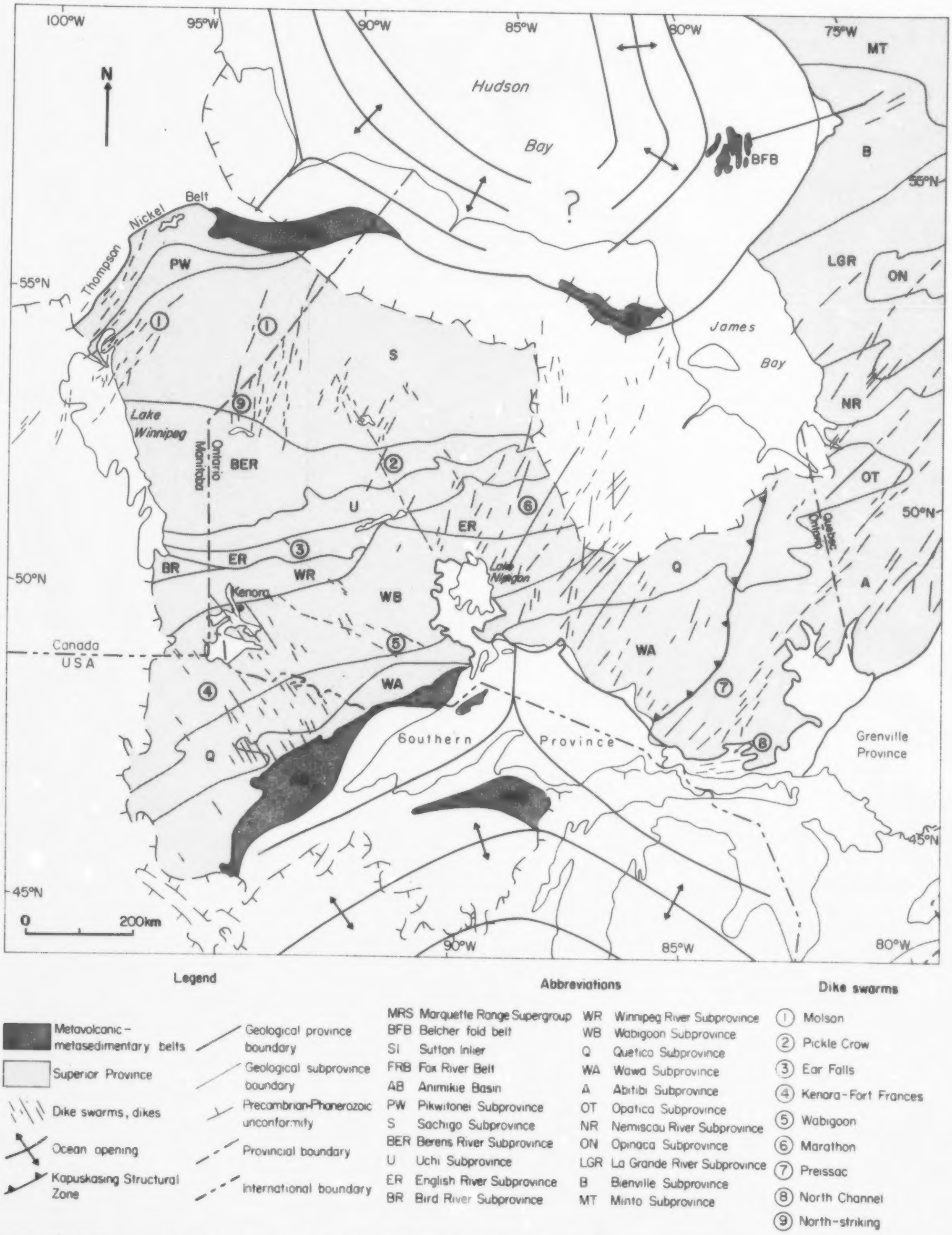
The Kenora–Fort Frances dikes are generally mildly to highly altered (Southwick and Day 1983; Hanson and Malhotra 1971) and are composed of variable amounts of plagioclase, pyroxene, quartz, hornblende and opaques. The secondary minerals include blue-green hornblende, chlorite, actinolite, calcite, epidote and biotite (Southwick and Day 1983). The texture of the dikes is hypidiomorphic granular, with grain sizes up to 8 mm.

Some of the larger gabbroic dikes show composite intrusions with a central dioritic zone and an outer zone of gabbro with chilled diabasic contacts with the country rock. The transition zone (up to 5 m) consists of alternating pyroxene-rich and plagioclase-rich layers 1 to 5 cm thick (Ayer and Buck 1989). Southwick and Day (1983) postulated that late-stage fractionation occurred as indicated by dioritic central zones in many dikes and rare examples of late evolved diorite liquid in already crystallized gabbroic dikes.

The Kenora–Fort Frances dikes occupy northwest-oriented fractures that are apparently unrelated to the main Archean deformation which affected the greenstone belts (Southwick and Day 1983; Ayer and Buck 1989). The dikes are overlain by sedimentary rocks of the Animikie Group (Southwick and Day 1983). Post-emplacement movement has been recorded by Southwick and Day (1983) near the Mesabi Iron Range, Minnesota.

There have been numerous interpretations of the tectonic setting of the Kenora–Fort Frances dike swarm. Fahrig (1987) interpreted the dike swarm as a possible failed arm associated with the formation of the Animikie Basin. Rubidium-strontium and K-Ar dates from dikes of the Kenora–Fort Frances swarm suggest that crustal extension and associated magmatism probably occurred between 2120 and 2240 Ma (Hanson and Malhotra 1971; Southwick and Day 1983). Downwarping, rifting and sedimentation probably continued until the onset of the compressional stages of the Penokean Orogeny approximately 1850 Ma.

Hoffman (1987), and Southwick and Halls (1987), however, proposed an alternate model. They suggested that the Animikie Basin is a foredeep basin that developed in front of the advancing Penokean fold and thrust belt. In this model, Southwick and Halls (1987) suggest that the Kenora–Fort Frances dike swarm may be related to peripheral bulging as a consequence of thrust loading of continental crust, and extension that occurred on the cratonic side of a Proterozoic flexural foredeep basin. Although the radial pattern of the Kenora–Fort Frances dike swarm with respect to the Penokean Orogen is consistent with this model, precise dating is required to show that the age of the igneous


100°W
95°W
90°W
85°W
80°W
75°W
N
Hudson
Bay
?
BFB
MT
B
55°N
LGR
ON
Belt
Nickel
Thompson
PW
55°N
①
①
S
James
Bay
NR
⑨
Lake
Winnipeg
Ontario
Manitoba
BER
②
OT
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U
ER
Quebec
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③
BR
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Lake
Nipigon
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50°N
Kenora
WB
Canada
USA
⑤
WA
④
WA
⑦
Southern
Province
⑧
Grenville
Province
Q
45°N
45°N
0
200km
90°W
85°W
80°W
Legend
Metavolcanic – metasedimentary belts
Superior Province
Dike swarms, dikes
Ocean opening
Kapuskasing Structural Zone
Geological province boundary
Geological subprovince boundary
Precambrian-Phanerozoic unconformity
Provincial boundary
International boundary
Abbreviations
MRS Marquette Range Supergroup
BFB Belcher fold belt
SI Sutton Inlier
FRB Fox River Belt
AB Animikie Basin
PW Pikwitonei Subprovince
S Sachigo Subprovince
BER Berens River Subprovince
U Uchi Subprovince
ER English River Subprovince
BR Bird River Subprovince
WR Winnipeg River Subprovince
WB Wabigoon Subprovince
Q Quetico Subprovince
WA Wawa Subprovince
A Abitibi Subprovince
OT Opatica Subprovince
NR Nemiscau River Subprovince
ON Opinaca Subprovince
LGR La Grande River Subprovince
B Bienville Subprovince
MT Minto Subprovince
Dike swarms
① Molson
② Pickle Crow
③ Ear Falls
④ Kenora-Fort Frances
⑤ Wabigoon
⑥ Marathon
⑦ Preissac
⑧ North Channel
⑨ North-striking

Figure 17.4. Paleoproterozoic (2.17 to 1.88 Ga) dikes and related metavolcanic-metasedimentary belts on the north and south perimeter of the Superior Province. The origin of the dike swarms and metavolcanic-metasedimentary belts are probably related to ocean openings and subsequent closings.

rocks is similar to that of the Penokean Orogen (Southwick and Halls 1987).

Nipissing Diabase and Preissac Dike Swarm

The $2219^{+3.6}_{-3.5}$ million-year-old (Corfu and Andrews 1986) Nipissing diabase occurs over a wide area between Sault Ste. Marie in the west and Cobalt in the east.

The Nipissing diabase intrudes Paleoproterozoic Huronian metasediments and consists mainly of pyroxene gabbro and hornblende gabbro, with minor amounts of feldspathic pyroxenite, gabbroic anorthosite and granophyre (Choudhry 1984; Buchan et al. 1989). These rocks form mafic dikes, sills and cone sheets. Although olivine is present in some intrusions, the typical minerals of the Nipissing diabase are: clinopyroxene, orthopyroxene, plagioclase and hornblende with little quartz, ilmenite-magnetite and sulphides; and minor amounts of secondary amphibole, biotite, chlorite, apatite and epidote (Choudhry 1984; Buchan et al. 1989).

Dikes of the Nipissing diabase strike north to northeast and range in width from a few metres to several hundred metres. Emplacement of the Nipissing dikes appears to be largely controlled by pre-existing structures in the Huronian and Archean rocks (Choudhry 1984). Both sills and dikes occur in the same general area as the Huronian metasediments. The dikes are generally a few metres wide and northeast striking (Choudhry 1984), whereas sills are up to several hundred metres thick. The Nipissing dike swarm has a restricted areal extent. Based on paleomagnetic evidence, the Nipissing diabase sheets represent 3 distinct intrusive events which are thought to have occurred between 2.15 and 2.25 Ga (Buchan et al. 1989). Buchan et al. (1989) have correlated the N_2 remanence direction (2.15 Ga) of Nipissing diabase with the northeast-striking Preissac dikes.

The northeast-striking Preissac dike swarm is a 2.15 billion-year-old (Gates and Hurley 1973; Hanes and York 1979) quartz diabase swarm. The swarm is petrographically and geochronologically similar to the Nipissing diabase (Fahrig and West 1986). The age of the Preissac dikes to the west of the Kapuskasing Structural Zone is uncertain, since most of the radiometric and paleomagnetic ages were determined for dikes situated south and southeast of the Kapuskasing Structural Zone. The dikes on either side of the Kapuskasing Structural Zone are correlated on the basis of strike, petrochemistry (Thurston et al. 1977; Ernst 1981; Condie et al. 1987) and paleomagnetic direction (Irving and Naldrett 1977; Ernst 1981; Halls and Shaw 1987). Further west, along the northern shore of Lake Superior, the Preissac dikes are not readily distinguished from those of the Marathon dike swarm (2.17 Ga)(Fahrig and West 1986; Fahrig 1987).

The Preissac dikes are several hundred kilometres long, up to 300 m wide, and have subvertical to vertical dips. They are usually composed of plagioclase (50 to 60%), augite (30%), quartz (5 to 10%) and opaques (5 to 7%). The plagioclase occurs as stubby to acicular laths up to 2 mm long and bears an ophitic to subophitic relationship to the enclosing augite, which is generally altered to chlorite, amphiboles and biotite (Ernst 1981).

The Preissac dikes cut obliquely across the structural grain of the Archean and Paleoproterozoic host rocks. They generally show fresh, unaltered margins, with lower greenschist- to subgreenschist-grade deuteric and/or hydrothermal alteration in the interiors (Ernst 1981). Most dikes are subparallel to, or occupy, the regional northeast-striking faults, as observed by the author (unpublished data) and R. P. Sage (Ontario Geological Survey, personal communication) in Wawa and surrounding areas. The ages of the faults are not known; however, the fresh, unaltered dikes within, and along, these faults suggest that the faults probably predate the Preissac dike swarm. Preissac dikes to the southeast of Kapuskasing Structural Zone are cut by some of the younger dikes (e.g., 1.14 billion-year-old Great Abitibi Dike, Ernst 1989), and by numerous northwest-striking faults.

The direct correlation of the Preissac dike swarm with a specific tectono-magmatic event is not clear and may require more precise dating of the swarm than currently exists. As suggested by Fahrig (1987), the Preissac dike swarm, like the Kenora–Fort Frances and the Marathon dike swarms, may be related to tectonic events in the Animikie Basin and Penokean Orogen to the south (*see* Figure 17.4).

Marathon Dike Swarm and Sutton Inlier Diabase Sills and Dikes

Dikes of the Marathon swarm strike northeasterly and lie to the north of Lake Superior (*see* Figures 17.1 and 17.4). The extent of this swarm is not well constrained, largely because the swarm is not clearly distinguishable from other dike swarms of various trends and ages. A few dikes can be traced aeromagnetically for several hundred kilometres, under Paleozoic cover, from Lake Superior to the James Bay area.

Very little published work is available on the Marathon swarm. The paleomagnetic work of Fahrig et al. (1965) provided a pole for the dike swarm which Fahrig (1987) correlated to the pole obtained by Halls (1986) for the Kenora–Fort Frances dike swarm. Based on this correlation, Fahrig (1987) and Fahrig and West (1986) suggested an age of 2.17 Ga for the Marathon dike swarm. A K-Ar age of 1.96 Ga has also been reported (Wanless et al. 1966) from an olivine diabase dike in the Miminiska Lake area.

Diabase-gabbro bodies, which intrude sediments of the Sutton Inlier, are exposed along a northwest-trending belt from east of Nawashe Lake to Sutton Ridges (Bostock 1971). Typically, the diabase is fine to medium grained, but coarse-grained and porphyritic varieties are also common. The diabase is usually exposed in the form of ridges and hills with prominently developed columnar jointing. The diabase sheets, measuring 10 to 38 m thick, occur in the Aquatuk and Sutton River areas (Bostock 1971). Two diabase dikes, measuring approximately 1 m and 15 m thick, are reported to intrude metasediments of the Sutton Ridges and Nawashe Lake, respectively (Bostock 1971). The diabase sheets and dikes are characterized by the presence of, in order of decreasing abundance, plagioclase, pyroxene, chlorite, opaques (magnetite-ilmenite and iron sulphides) and apatite (Bostock 1971). Minor amounts of secondary amphibole, sericite, epidote and carbonates occur in altered diabase.

Based on geophysical data, the igneous and sedimentary rocks of the Sutton Inlier have been correlated with the Belcher fold belt in the eastern Hudson Bay and with the Fox River Belt in Manitoba (Bostock 1971; Ricketts and Donaldson 1981; Hoffman 1988). These belts are interpreted to form part of the 1.9 to 1.8 billion-year-old Trans–Hudson Orogen (Hoffman 1988).

The Marathon dike swarm, along with the Preissac and Kenora–Fort Frances dikes, are related, by Fahrig (1987), to events in the Animikie Basin and Penokean Orogen. However, as the dikes of the swarm also strike directly toward the Trans–Hudson Orogen (1.9 to 1.81 Ga, Hoffman 1988), which is situated along the northern perimeter of the Superior Province, mechanical principles suggest an alternate hypothesis that some dikes might be related to the orogen (*see* Figure 17.4).

The igneous rocks within the Belcher fold belt are thought to have originated from fissures during the development of Richmond–Belcher rift in Paleoproterozoic time (2.0 Ga, Ricketts and Donaldson 1981) and it is possible that the emplacement of the Marathon dike swarm may have been part of this rifting stage.

Molson Dike Swarm and North-Striking Dikes

The mafic to ultramafic Molson dike swarm, dated at 1883±2 Ma (Heaman et al. 1986), derives its name from Molson Lake in Manitoba (Fahrig et al. 1965). The dikes of the swarm strike northeasterly (10° to 60°) and display near-vertical dips (*see* Figure 17.4). They cut obliquely across the structural trend of the gneisses of the Pikwitonei region and supracrustal sequences within the northwestern Superior Province of Manitoba and Ontario. The dike swarm appears to be bounded by the Split Lake fault and the Fox River Belt to the north, and it does not extend south of latitude 52°N, or west of the Thompson Nickel Belt (Scoates and Macek 1978). The eastern limit of the dike swarm is uncertain. Molson dikes occur as far east as North Caribou Lake in northwestern Ontario; therefore, the eastern limit of the dike swarm is tentatively placed there (*see* Figure 17.4). The concentration of dikes is considerably greater in the west, toward the Thompson Nickel Belt, than east of 97°W longitude. This indicates that the tensional forces giving rise to the crustal extension, were greatest in that area (Scoates and Macek 1978).

The largest dike of the Molson swarm in northwestern Ontario, occurs near the Ontario–Manitoba border (*see* Figure 17.4). It extends north-northeasterly for approximately 360 km, from south of Favourable Lake, Ontario, to just south of the Fox River Belt in Manitoba. This dike, known as the "Kistigan Lake dike" (Scoates and Macek 1978), is referred to here as the "Favourable Lake–Kistigan Lake dike". The dike has been mapped by Ayres (1974) and Wilson (1987) in the Trout Lake and Lingman Lake areas, respectively. The Favourable Lake–Kistigan Lake dike is a north-northeasterly (10° to 20°) striking, steeply to vertically dipping dike which ranges in width from 15 to 70 m. Several subsidiary dikes, ranging in width from a few millimetres to up to 10 m, and also generally striking to the north-northeast, are common near the main dike (Ayres 1974). The dike has a fine-grained chilled margin and a medium-grained central zone (Ayres 1974; Wilson 1987). The chilled margin, up to 10 cm wide, consists of serpentine pseudomorphs after olivine, pyroxene and acicular plagioclase. These minerals are always oriented parallel to the contact (Ayres 1974). The central medium-grained zone is composed of plagioclase (40 to 60%), pyroxene (30 to 47%), olivine (7 to 15%) and opaques (up to 20%) (Ayres 1974; Wilson 1987).

Recent geological mapping (Stone 1989) and aeromagnetic surveys (*see* Map 2584, map case) have revealed several north-striking mafic dikes in the Favourable, Sandy and Lingman lakes areas. Petrographic data (D. Stone, unpublished data) indicate that north-striking dikes are chiefly composed of plagioclase, augite, tremolite, sericite and opaques, and contain no olivine; whereas the Molson dikes typically consist of olivine, pyroxene, plagioclase and opaques. The north-striking dikes are generally narrow (less than 10 m), which is reflected on the aeromagnetic maps by their weak magnetic signature, and contrasts with the 10 to 70 m wide Molson dikes. The age range and geochemistry of these north-striking dikes, and the tectono-magmatic event with which they can be correlated, are presently unknown; therefore they are grouped with the Molson dike swarm until more details are known about them.

The fold belts that border the western and northern perimeter of the Superior Province are exposed in 4 segments from west to east, the Thompson Nickel, Fox River, Belcher and Cape Smith belts. These belts have been dated at 2.0 to 1.8 Ga (Hoffman 1988), and their igneous and sedimentary rock sequences provide a record of early rifting followed by a collisional orogen at 1.91 to 1.81 Ga (Hoffman 1988).

The east-trending Fox River Belt (Scoates and Macek 1978; Fahrig 1987) comprises steeply north-dipping thrust stacks of relatively thin basalt, shelf sediments and overlying foredeep flysch. The flysch contains differentiated tholeiitic-komatiitic sills and flows (Hoffman 1988), including the Fox River sill. The Fox River sill and the Molson dike swarm have similar chemistry (Scoates and Macek 1978) and ages (1883±2 Ma, Heaman et al. 1986). The Molson dike swarm is considered to represent a failed-arm dike set related to development of the Fox River Belt (Fahrig 1987).

Sudbury Dike Swarm

The 1238±4 million-year-old (Krogh et al. 1987) Sudbury dike swarm trends west-northwesterly, and extends northward from the southwestern part of the Grenville Front Tectonic Zone (Figure 17.5). The northwestern limit of the swarm is not known, largely because the dike swarm is intermixed with the Matachewan and Hearst, and Keweenawan dike swarms which have similar trends. The Sudbury dikes, generally 15 to 30 m thick, with a few up to 100 m, are most abundant in the Sudbury area. They dip vertically, weather

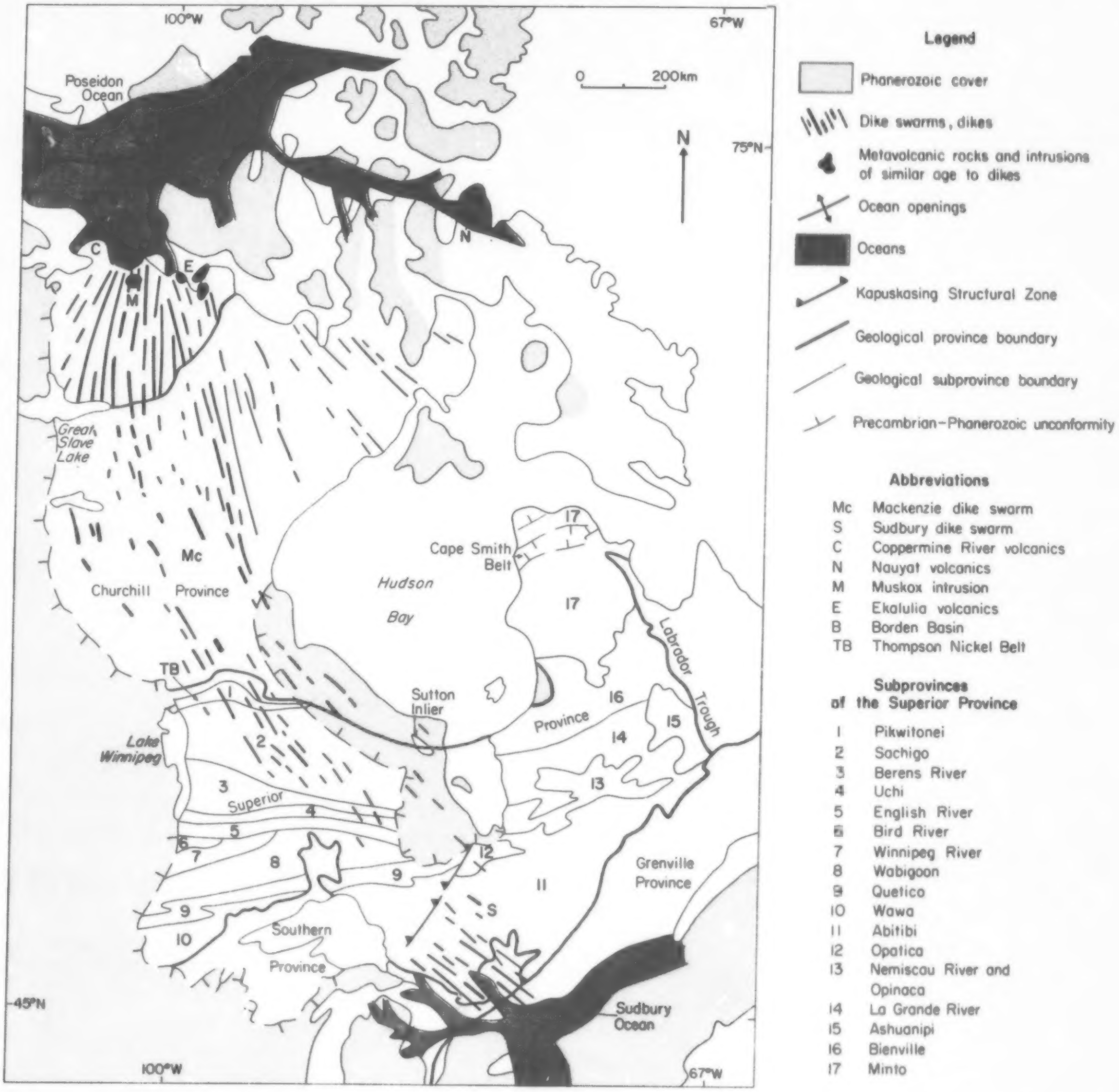
100°W
67°W
Poseidon Ocean
0 200km
N
75°N
C
E
M
N
Great Slave Lake
Mc
Churchill Province
TB
Cape Smith Belt
Hudson Bay
17
16
15
14
13
Labrador Trough
Sutton Inlier
Province
Lake Winnipeg
1
2
3
Superior
4
5
6
7
8
9
10
11
12
S
Southern Province
Grenville Province
Sudbury Ocean
45°N
Legend
Phanerozoic cover
Dike swarms, dikes
Metavolcanic rocks and intrusions of similar age to dikes
Ocean openings
Oceans
Kapuskasing Structural Zone
Geological province boundary
Geological subprovince boundary
Precambrian–Phanerozoic unconformity
Abbreviations
Mc Mackenzie dike swarm
S Sudbury dike swarm
C Coppermine River volcanics
N Nauyat volcanics
M Muskox intrusion
E Ekalulia volcanics
B Borden Basin
TB Thompson Nickel Belt
Subprovinces of the Superior Province
1 Pikwitonei
2 Sachigo
3 Berens River
4 Uchi
5 English River
6 Bird River
7 Winnipeg River
8 Wabigoon
9 Quetico
10 Wawa
11 Abitibi
12 Opatica
13 Nemiscau River and Opinaca
14 La Grande River
15 Ashuanipi
16 Bienville
17 Minto

Figure 17.5. Mesoproterozoic (1.2 Ga) Mackenzie and Sudbury dike swarms, and their suggested zones of spreading (*modified from* Fahrig 1987).

recessively, and are consequently generally marked by narrow, linear valleys (Card and Innes 1981).

The majority of the Sudbury dikes are olivine tholeiites, and composed of plagioclase (60%), olivine (15%), titaniferous augite (12%), magnetite-ilmenite (5 to 10%); with minor amounts of chlorite and biotite (Card and Innes 1981; Choudhry 1984; Gibson et al. 1987). Most of these dikes are medium to coarse grained, with ophitic to subophitic textures (Card and Innes 1981; Choudhry 1984).

The Sudbury dikes transect lithotectonic trends in Archean and early Paleoproterozoic rocks, but are displaced by faults of the Grenville Front Tectonic Zone, hence they are younger than orogenic events in the Southern Province, but older than terminal stage docking of the Grenville Province against the Superior Province. The Sudbury dike swarm appears to converge in trend towards a common point to the southeast; therefore, Fahrig (1987) suggested that the swarm may be ascribed to a zone of spreading related to a hypothetical "Sudbury ocean" to the southeast. In an alternate explanation, Davidson and Bethune (1988) have suggested that the swarm may be related to the Grenville collision.

Mackenzie Dike Swarm

The northwesterly striking Mackenzie dikes (1267±2 Ma, LeCheminant and Heaman 1989) constitute the largest Precambrian dike swarm in the world. The dike swarm fans out southward in a radial fashion for approximately 2400 km (LeCheminant and Heaman 1989) from the Arctic coast near Coronation Gulf, to northwestern Ontario (*see* Figure 17.5). The widest part of the dike swarm is approximately 1800 km across (Fahrig and West 1986). The dikes are most abundant in the Slave Province. A small fraction of the Mackenzie dike swarm has been recognized in the extreme northwestern part of Ontario.

Very little is known about the petrography, chemistry or physical characteristics and ages of these dikes in Ontario. The correlation of dikes in northwestern Ontario with well-studied examples in the Slave Province is mainly based on the similarities in strike, petrography and aeromagnetic characteristics. The following brief description of the petrography and physical properties of these dikes is excerpted from studies in Ontario by Ayres (1969), Bennett and Riley (1969), Thurston et al. (1979), Gibson et al. (1987) and other workers who have studied these dikes outside of Ontario.

The Mackenzie dikes are generally composed of plagioclase, clinopyroxenes, biotite, opaques and quartz. Accessory minerals include chlorite, amphiboles, and trace amounts of apatite. The dikes are usually 30 to 60 m wide and dip subvertically. They are characterized by distinct, linear, positive magnetic anomalies. Dikes greater than 15 m wide are easily detected on the 1 inch to 4 miles (1:253 440) federal and provincial aeromagnetic maps. Generally, thin dikes (less than 15 m) have either very weak linear anomalies or no magnetic expressions. A recent high resolution aeromagnetic survey (OGS 1985) and mapping by Bartlett et al. (1985) and Breaks et al. (1987a, 1987b), in the North Caribou Lake area, identified several new dikes of this swarm of variable widths and lengths. One dike, 10 to 20 m wide, is situated just north of Eyapamikama Lake and extends for approximately 150 km to just southeast of Lingman Lake.

The Mackenzie dike swarm, within the Sachigo Subprovince of northwestern Ontario, crosscuts the west- and northwest-trending structural grain of the Archean host rocks. A few are parallel to, or occupy, regional west-northwest-oriented faults and shear zones, which were produced during the late Archean by a north-northwest-directed compressional event (Osmani and Stott 1988; Osmani et al. 1989). On the basis of crosscutting relations between the dikes and the regional structures, it is concluded that the emplacement of the Mackenzie dike swarm was not greatly affected by Archean structural fabric.

The Mackenzie dikes are terminated along the Arctic coastline, where a series of basins are located containing 1.3 billion-year-old volcanic rocks. These basins are largely rift controlled. The 1.26 billion-year-old Muskox intrusion and correlative Coppermine River volcanic rocks reflect rifting and the development of a passive-margin sequence related to the opening of the Poseidon ocean Fahrig 1987; (LeCheminant and Heaman 1989). The Mackenzie dike swarm is spatially related to these rocks, and is probably also a manifestation of this event.

Abitibi Dike Swarm

The east-northeast-trending Abitibi dike swarm in southeastern Superior Province consists of at least a dozen major, and several smaller, dikes (Figure 17.6). The longest of these dikes is the Great Abitibi Dike (Ernst 1982, 1989; Ernst et al. 1987) which can be traced for approximately 600 km, extending from near Foleyet, Ontario, to Chibougamau, Quebec. It essentially crosscuts the Abitibi Subprovince from the Kapuskasing Structural Zone to the Grenville Front Tectonic Zone. The Great Abitibi Dike is up to 250 m wide and subvertical in attitude. The age of this dike, and probably the rest of the swarm, is 1140.6±2 Ma (Krogh et al. 1987). Other than the Great Abitibi Dike and its subsidiary dikes in the Timmins–Kirkland Lake area, very little is known about the dikes of Abitibi swarm, especially those west of the Kapuskasing Structural Zone. The latter dikes are correlated with the Great Abitibi Dike on the basis of petrography, strike, and to some extent, on the basis of crosscutting relationships with dike swarms of known ages.

The Great Abitibi Dike is composed of unaltered olivine gabbro to monzodiorite and consists of plagioclase (40 to 70%), alkali feldspar (up to 15%), titaniferous clinopyroxene (8 to 25%), olivine (up to 20%), opaques (5 to 15%), apatite (up to 5%) and biotite (up to 5%); minor quartz and orthopyroxene are rarely present (Ernst et al. 1987). In the Kabenung and Mishibishu lakes areas, Sage (in prep.) reports a dike up to 300 m wide, traceable from the north shore of Lake Superior in the southwest, to well beyond Paleozoic cover in the northeast. This dike is probably part of the Abitibi dike swarm. A few smaller dikes of similar strike and magnetic intensity, west and east of this dike, are also interpreted to be part of the Abitibi dike swarm.

The Abitibi dike swarm is related to the Keweenawan volcanic rocks of the Superior basin on the basis of age and geochemistry. The radiometric age (1.14 Ga) and paleomagnetic signature of the Abitibi dike swarm suggests emplacement during a lower "normal" polarity interval of Lake Superior Keweenawan stratigraphy (Ernst and Bell 1987). The Abitibi dikes appear to fan out from the Superior basin in a manner that suggests radiation from a centre of rifting (*see* Figure 17.6).

Keweenawan Dike Swarms

Three dike swarms that are exposed along the north shore of Lake Superior are named, from northwest to northeast, the Pigeon River, Nipigon and Pukaskwa dike swarms. Their locations in relation to the Midcontinent Rift are shown in Figure 17.6. These dike swarms are related to volcanic rocks of the Keweenawan Supergroup that were deposited in the rift during the interval 1109^{+4}_{-2} Ma (Davis and Sutcliffe 1985) to $1086^{+1.3}_{-3.0}$ Ma (Palmer and Davis 1987).

The Pukaskwa dikes strike northwesterly along the northeast shore of Lake Superior (*see* Figure 17.6). The dike swarm extends from Batchawana Bay area to Marathon. Structurally, the Pukaskwa dikes dip at high angles (50°NE) to the Midcontinent Rift, and follow the orientation of the rift (Green et al. 1987; Halls and Shaw 1987).

The dikes are composed of olivine diabase, and consist of plagioclase (45%), iddingsite after olivine (10 to 20%), olivine (1 to 5%) and augite (20%), with the remainder being mainly iron-titanium oxides with minor chlorite (Thurston et al. 1977). The relatively high magnetite content of these dikes is reflected by the strong aeromagnetic signature in the University River area. The dikes intrude Archean rocks and cut the northeast-striking quartz diabase dikes (Preissac dikes?).

The Pigeon River dike swarm occurs along the northwestern shore of Lake Superior (*see* Figure 17.6) in the vicinity of the Ontario–Minnesota border, and extends east to Black Bay Peninsula. The majority of the dikes strike northeasterly, and are roughly parallel to the Midcontinent Rift. A few dikes strike northwesterly and are normal to the rift. The northeast-striking dikes dip steeply to the southeast. The individual dikes are up to 40 km long and greater than 100 m wide, and a few dikes attain widths up to 500 m (Green et al. 1987). The Pigeon River dikes cut the lower Paleoproterozoic Rove Formation, and early Keweenawan igneous units such as the Logan sills (Geul 1970; 1973) and Osler Formation (McIlwaine and Wallace 1976). The Pigeon River dikes are composed of plagioclase, clinopyroxene, olivine and magnetite, and are typically aphyric with ophitic clinopyroxene (Smith and Sutcliffe 1987). The Pigeon River dikes have both normal and reversed magnetic polarity (Pesonen 1979), and probably were feeders for the olivine tholeiite basalts of the Keweenawan Supergroup.

Dikes situated along the north-central shore of Lake Superior, and striking in a northerly direction along the east

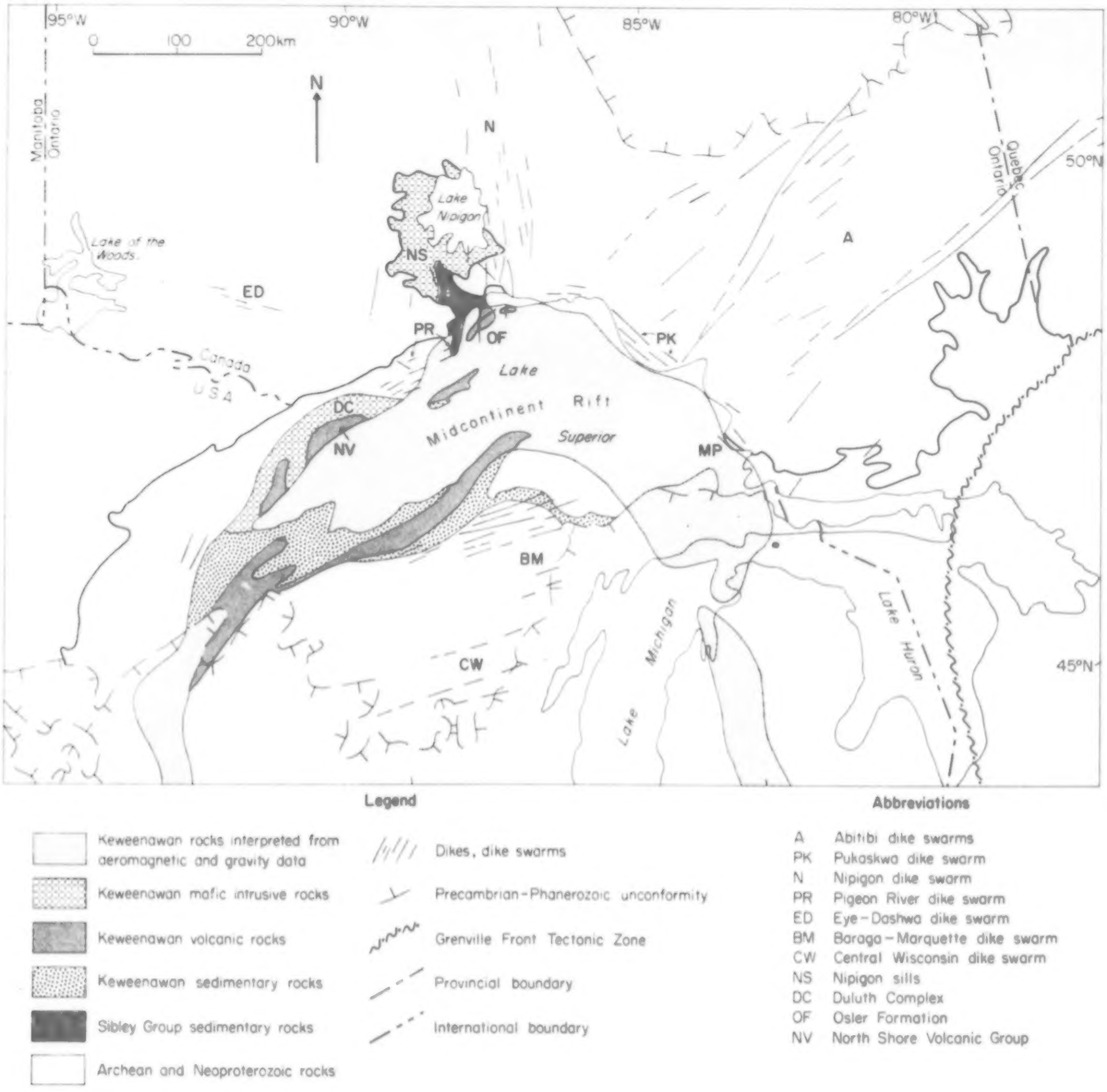
95°W
90°W
85°W
80°W
50°N
45°N
0
100
200km
N
Manitoba
Ontario
Quebec
Ontario
Lake Nipigon
Lake of the Woods
Canada
U.S.A.
Lake Superior
Midcontinent Rift
Lake Michigan
Lake Huron
N
NS
PR
OF
ED
PK
A
DC
NV
MP
BM
CW
Legend
Keweenawan rocks interpreted from aeromagnetic and gravity data
Keweenawan mafic intrusive rocks
Keweenawan volcanic rocks
Keweenawan sedimentary rocks
Sibley Group sedimentary rocks
Archean and Neoproterozoic rocks
Dikes, dike swarms
Precambrian–Phanerozoic unconformity
Grenville Front Tectonic Zone
Provincial boundary
International boundary
Abbreviations
A Abitibi dike swarms
PK Pukaskwa dike swarm
N Nipigon dike swarm
PR Pigeon River dike swarm
ED Eye–Dashwa dike swarm
BM Baraga–Marquette dike swarm
CW Central Wisconsin dike swarm
NS Nipigon sills
DC Duluth Complex
OF Osler Formation
NV North Shore Volcanic Group

Figure 17.6. Generalized geological map of the Lake Superior area, showing location of mafic dike swarms related to the Midcontinent Rift.

Table 17.2. Major and trace element averages of some of the Proterozoic mafic dike swarms.

	Matachewan and Hearst	Preissac (NE)	Kenora–Fort Frances	Molson	Mackenzie	Sudbury	Abitibi	Keweenawan (Nipigon)	
								Olivine diabase	Quartz diabase
SiO_2	51.0	50.4	50.0	49.4	49.4	47.4	47.3	48.0	49.0
TiO_2	1.4	1.4	1.7	0.9	2.5	2.9	3.4	1.6	1.5
Al_2O_3	13.6	13.6	13.0	13.5	13.2	15.3	17.7	14.3	14.4
FeO^t	14.12	12.61	16.29	10.48	13.98	14.24	12.22	12.16	13.14
MnO	0.22	0.22	0.21	0.18	0.21	0.20	0.23	0.22	0.20
MgO	5.6	5.5	6.2	9.0	6.0	5.5	5.3	5.7	6.0
CaO	9.5	10.2	10.0	11.0	9.4	7.4	8.8	9.5	9.9
Na_2O	2.3	2.2	2.0	1.6	2.0	3.3	3.4	2.8	2.1
K_2O	0.8	0.5	0.4	0.4	0.8	1.4	1.3	1.4	0.9
P_2O_5	0.17	0.14	0.14	0.10	0.28	0.66	0.81	0.60	0.20
Mg Number	44	46	43	63	46	43	46	48	47
Ba	209.0	199.0	—	63.0	267.0	496	840.0	797	218
Rb	28.0	22.0	17.5	20.0	13.5	41.0	13.0	53.0	—
Th	2.3	1.9	—	—	2.3	2.5	1.3	—	—
Nb	10.0	11.0	—	—	10.0	16.0	15.0	19.0	18.0
Ta	0.6	0.6	—	—	1.0	0.9	1.4	—	—
La	13.5	12.5	—	—	25.5	27.0	24.0	—	—
Ce	28.3	27.5	—	—	46.0	60.0	94.0	—	—
Sr	179.0	181.0	88.0	90.0	182.0	439.0	882.0	485.0	142.0
Nd	—	—	—	—	28.0	35.0	—	—	—
Sm	4.4	3.9	—	—	6.3	7.0	9.8	—	—
Zr	122.0	99.0	87.0	97.0	118.0	192.0	108.0	172.0	88.0
Hf	3.4	2.8	—	—	4.3	5.0	5.0	—	—
Eu	1.5	1.4	—	—	2.0	2.6	3.8	—	—
Cs	1.3	1.7	—	—	—	1.0	1.0	—	—
V	352.0	354.0	—	310	327.0	214.0	235.0	287.0	297.0
Tb	0.9	0.8	—	—	0.9	1.2	1.3	—	—
Y	37.4	29.0	34.0	—	28.5	36.0	25.0	26	22
Yb	3.2	3.0	—	—	2.2	3.4	4.0	—	—
Samples	30	6	14	22	7	11	121	7	6
References	Condie et al. (1987) Nelson et al. (1990)	Condie et al. (1987)	Southwick and Halls (1987)	Ermanovics and Fahrig (1975)	Gibson et al. (1987)	Condie et al. (1987) Gibson et al. (1987)	Condie et al. (1987) Ernst et al. (1987)	Kresz and Zayachivsky (1988) Carter (1987) Mackasey and Wallace (1978)	

ᵗ*FeO + F_2O_3 expressed as FeO*

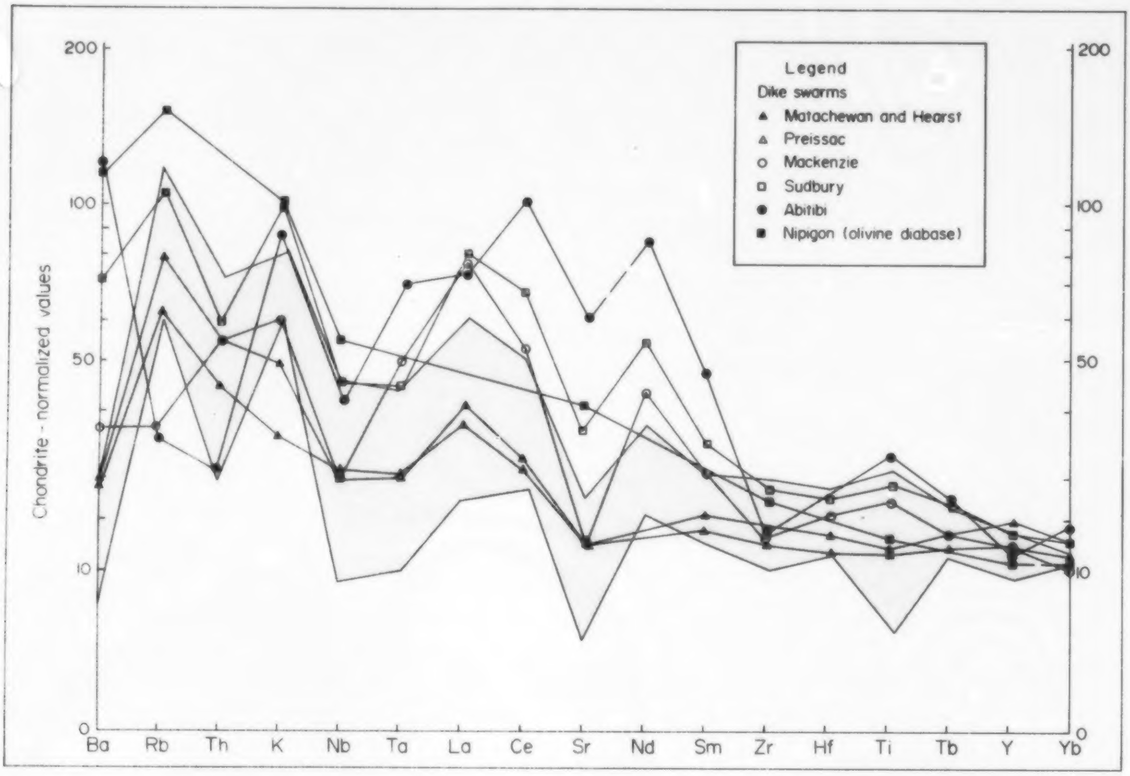

Figure 17.7.a) Chondrite-normalized trace element plots showing average composition of some Proterozoic dike swarms from the Superior Province. The normalization values are *from* Thompson et al. (1983). Shaded area represents the range in values for some Phanerozoic continental flood basalts in different parts of the world (Thompson et al. 1983).

shore of Lake Nipigon (*see* Figure 17.6), are referred to here as the Nipigon dike swarm. The extent of this dike swarm is difficult to assess because of the presence of dikes of various ages and strikes. The dikes are generally narrow; therefore, they are not readily recognized on regional aeromagnetic maps. Recent high resolution aeromagnetic coverage (OGS 1989) of the Beardmore–Geraldton Belt, east of Lake Nipigon, has resulted in the delineation of several new dikes.

The dike swarm has been relatively well studied immediately south and east of Lake Nipigon, compared to north and northeast of the lake. To the north and northeast, the dikes have been largely identified on the basis of strike and petrographic description (Amukun 1980; Thurston 1980; Wallace 1981).

Based on the recent studies of Sutcliffe (1986), Carter (1987) and Kresz and Zayachivsky (1988), the dikes range from a few metres to several kilometres in length, and up to 60 m wide. They generally have subvertical to vertical dips. The dikes range from olivine to quartz bearing in the mode and the norm. They are generally composed of calcic plagioclase, pyroxene (augite ± pigeonite), olivine, magnetite, with minor hornblende and biotite. Kresz and Zayachivsky (1988) reported aphyric and plagioclase-phyric types, and an olivine-pyroxene and plagioclase microporphyritic variety.

There is no radiometric age available on the Nipigon dike swarm. However, the dikes of the swarm intrude the Osler Formation along the north shore of Lake Superior, and the Nipigon sills (1.1 Ga, Davis and Sutcliffe 1985). Some of these dikes are also interpreted to have been the feeders of the Nipigon sills (Sutcliffe 1986).

Evolution of the Midcontinent Rift system at approximately 1.1 Ga is reviewed by Sutcliffe (this volume). Numerous intrusive bodies, including basaltic dike swarms, diabase sills and plutons of diverse shape and composition are associated with the volcanic rocks of the Lake Superior basin. The Keweenawan dike swarms, which represent the feeder system for volcanic rocks and diabase sills, occur peripheral to the basin (Green 1983; Sutcliffe 1986). The dikes strike both parallel to the basin axis and radiate outward. The volcanic equivalents of the Pukaskwa and Pigeon River dike swarms include the Osler Formation, Mamainse Point Formation and North Shore Volcanic Group which are situated along the northeast and northwest shores of Lake Superior, respectively. The dike swarms and the volcanic rocks provide a record of magmatism in an aborted rift (Cannon et al. 1989). In the Nipigon area, igneous rocks, including dikes which radiate north from the centre of the basin, are thought to represent a failed arm to the main rift in Lake Superior (Sutcliffe 1986).

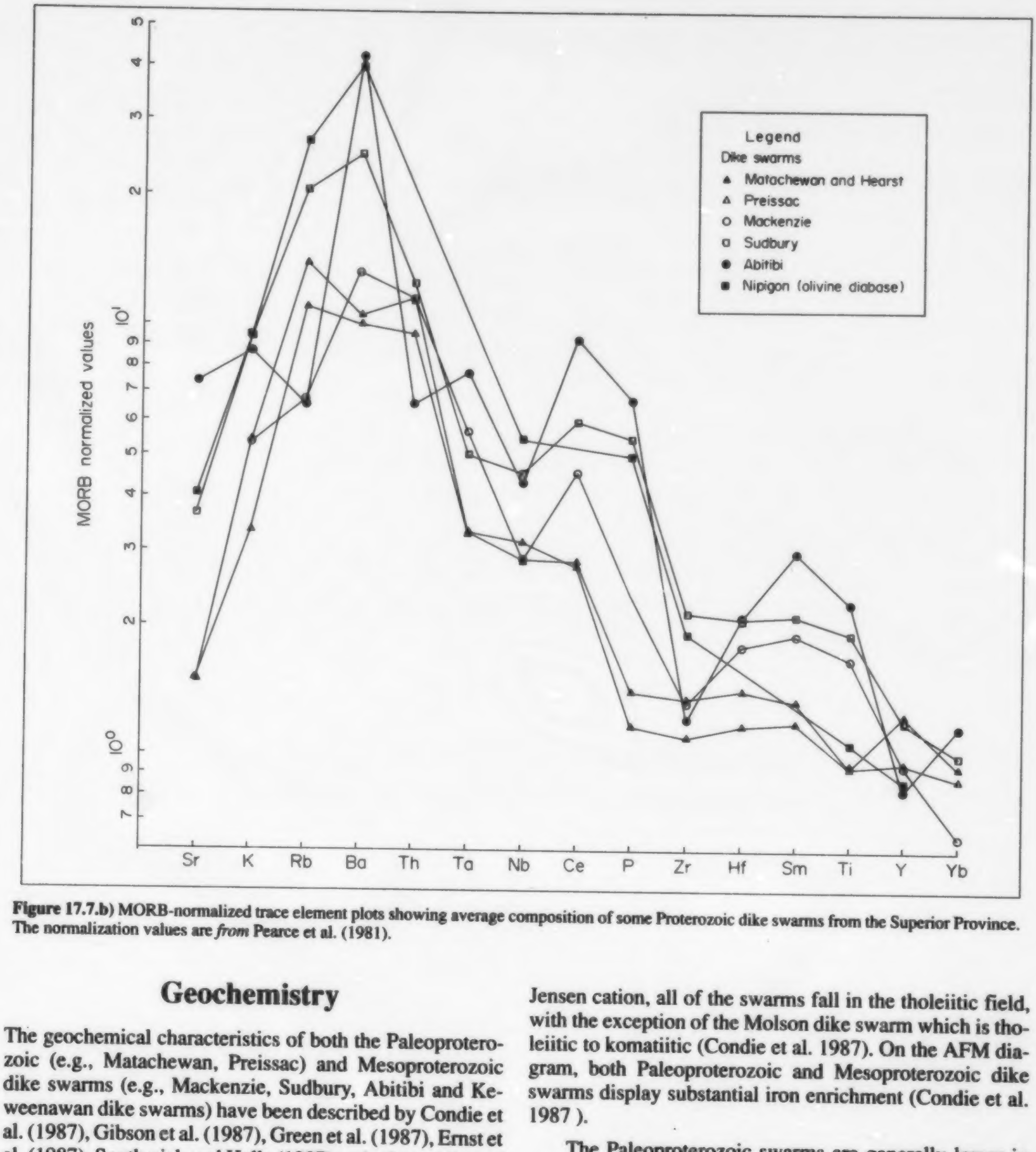

Figure 17.7.b) MORB-normalized trace element plots showing average composition of some Proterozoic dike swarms from the Superior Province. The normalization values are *from* Pearce et al. (1981).

Geochemistry

The geochemical characteristics of both the Paleoproterozoic (e.g., Matachewan, Preissac) and Mesoproterozoic dike swarms (e.g., Mackenzie, Sudbury, Abitibi and Keweenawan dike swarms) have been described by Condie et al. (1987), Gibson et al. (1987), Green et al. (1987), Ernst et al. (1987), Southwick and Halls (1987) and others. Most of the elemental (major and trace elements) averages given in Table 17.2 for mafic dike swarms of the Superior Province are compiled from the above published work. The interpretation of the data in the present study provides some generalizations regarding the magma source of dike swarms.

Proterozoic mafic dike swarms in the Superior Province of Ontario include both quartz (up to 4.9%) normative and olivine (up to 15%) normative diabase. Quartz diabase is the dominant composition in Paleoproterozoic dikes, whereas the majority of the Mesoproterozoic dikes are olivine diabase. On geochemical diagrams such as AFM and Jensen cation, all of the swarms fall in the tholeiitic field, with the exception of the Molson dike swarm which is tholeiitic to komatiitic (Condie et al. 1987). On the AFM diagram, both Paleoproterozoic and Mesoproterozoic dike swarms display substantial iron enrichment (Condie et al. 1987).

The Paleoproterozoic swarms are generally lower in Al_2O_3, TiO_2, P_2O_5 and alkalis compared to Mesoproterozoic swarms (*see* Table 17.2), which suggest that Mesoproterozoic swarms are slightly more evolved than Paleoproterozoic swarms. Figures 17.7a and 17.7b illustrate that Mesoproterozoic swarms are relatively enriched in most incompatible and slightly compatible elements compared to the Paleoproterozoic swarms. Both groups, however, display incompatible-element enrichment on both chondrite and MORB normalized spidergrams. The Mesoproterozoic Abitibi, Sudbury and Mackenzie dike swarms show slight peaks at Ti and prominent peaks at K, whereas the Paleoproterozoic Matachewan and Preissac dike swarms do not exhibit similar enrichment. All but Abitibi and Mackenzie dike swarms display peaks at Rb. Abitibi and Nipigon dikes show largest positive anomalies at Ba compared to all other dike swarms.

The Mesoproterozoic dikes show elevated abundances in most light rare earth elements (LREE)(30 to 100 times chondrite) compared to Paleoproterozoic dikes which range from 20 to 40 times chondrite values. The heavy rare earth element (HREE) distribution in both Paleoproterozoic and Mesoproterozoic swarms is nearly flat (10 to 20 times chondrites) (Figure 17.8). The Abitibi and Sudbury dike swarms show small positive Eu anomalies, indicating plagioclase accumulation.

The LREE enrichment of Mesoproterozoic dikes compared to Paleoproterozoic dikes suggest that Mesoproterozoic dike magmas were probably derived either from enriched-mantle sources or they represent contamination of mantle-derived basic magmas by sialic crust. The relatively less enriched Paleoproterozoic dike magma may represent a slightly depleted mantle source.

The prominent trough at Th, Nb, and Ta in Mesoproterozoic dikes is similar to that found in Phanerozoic flood basalts in many parts of the world, studied by Thompson et al. (1983)(*see* Figure 17.7a). In their study, the trough at Nb and Ta is attributed to crustal contamination as the primary cause for these negative anomalies.

In spite of the negative anomalies at Nb, Ta and Th, the Mesoproterozoic dike swarms in the Superior Province are relatively enriched in these elements and LREEs, which is a characteristic feature of many ocean island basalts (Thompson et al. 1983; Condie et al. 1987). On the other hand, the Paleoproterozoic swarms are relatively depleted in Nb, Ta and Th. Although Paleoproterozoic dikes show some resemblance to the continental flood basalts (*see* Figure 17.7a), they exhibit some trace-element (e.g., Ti and Zr, *see* Table 17.2) distribution more likely in common with calc-alkalic basalts (Condie et al. 1987).

It is also noteworthy that the consistent LREE enrichment pattern through Paleoproterozoic to Mesoproterozoic dikes is possibly the reflection of an increasing depth of melting in the mantle with decreasing age of dike swarms

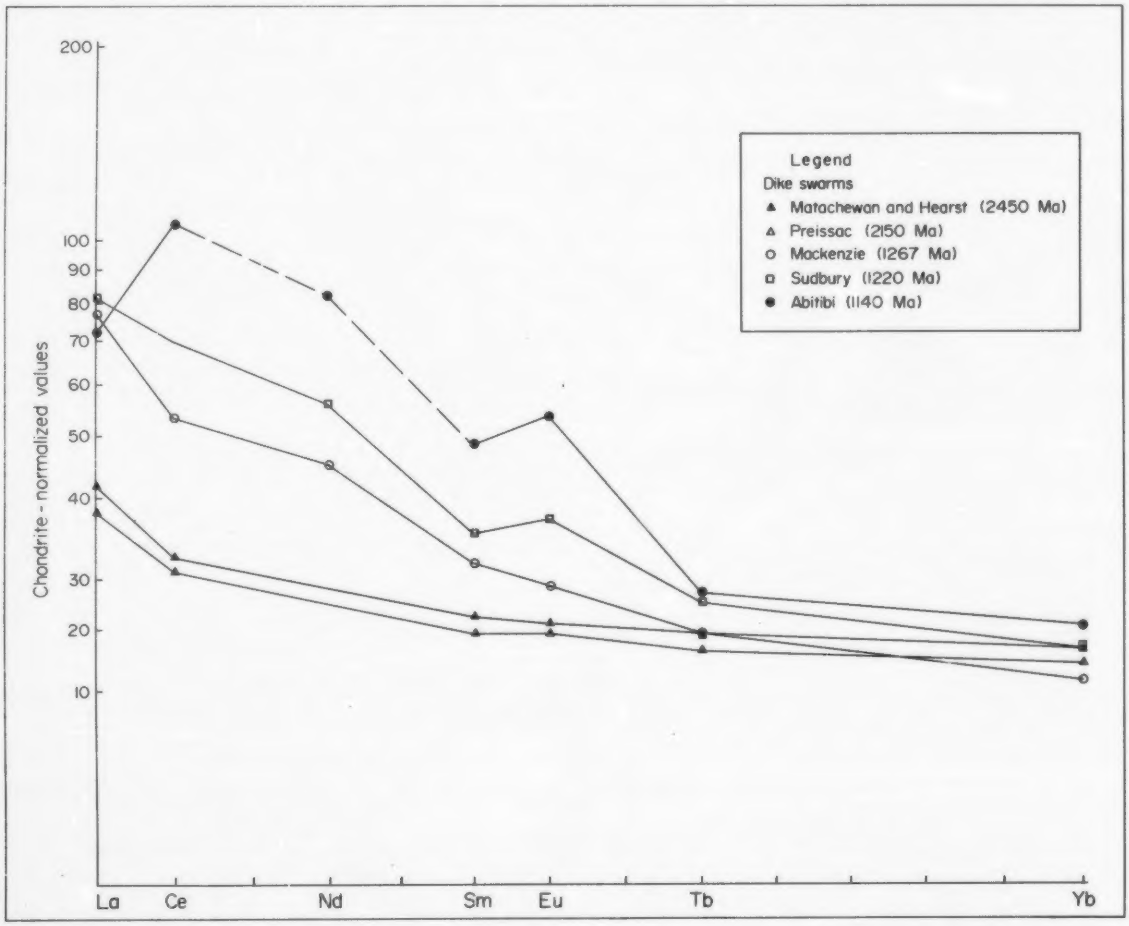

Figure 17.8. Chondrite-normalized REE distributions in average Proterozoic dike swarms from the Superior Province. Normalization values *from* Thompson et al. (1983).

(*see* Figure 17.8 and Table 17.2). The increased enriched mantle sources beneath the Superior Province in the Mesoproterozoic probably reflects increased recycling of sediments by subduction processes into the mantle after rapid continental growth in the Neoarchean (Condie et al. 1987).

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Chapter 18

Alkalic Rock, Carbonatite and Kimberlite Complexes of Ontario, Superior Province

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Precambrian Geology Section, Ontario Geological Survey

Abstract

The alkalic-carbonatite complexes and kimberlite intrusions of Ontario represent 5 intrusive events: 2.50 to 2.70 Ga, 1.70 to 1.85 Ga, 0.90 to 1.25 Ga, 450 to 650 Ma and approximately 150 Ma. Alkalic silicate rocks occur in relatively large stocks or ring complexes, and carbonatites occur in small plug-like bodies with diameters of 3 to 4 km. Kimberlites and kimberlitic rocks in the areas of Kirkland Lake and the Hudson Bay Lowland occur as small dikes and diatreme structures and are the most recent alkalic intrusive event. Since Archean time, the emplacement of alkalic rocks, carbonatites and kimberlites has had a close spatial relationship with regional crustal fractures such as the Trans-Superior Tectonic Zone, Kapuskasing Structural Zone and the Ottawa–Bonnechere Graben and its branches. Since multiple emplacement events occur in each fracture zone, these intrusions demonstrate that the Canadian Shield has undergone widespread periodic events involving metasomatism of depleted mantle, tectonism and alkaline magma emplacement.

The gabbroic phases of ring complexes host copper and platinum group element mineralization and the associated syenites have potential for building stone. Carbonatites contain reserves of niobium, uranium, phosphate and rare earth elements. Those complexes that have undergone karst-like weathering, concentrating the most desirable minerals, offer the greatest economic potential. Microdiamond has been found in Ontario kimberlite and additional exploration is warranted.

INTRODUCTION

The term "alkalic" is used to describe rocks that contain the alkali metals in concentrations greater than that commonly found in other igneous rocks. These alkali metal-rich compositions are reflected by the presence of alkali-rich and feldspathoid minerals in the rocks and are also accompanied by whole rock, volatile element, large ion lithophile (LIL), rare earth element (REE) and other incompatible element enrichment. In Ontario, the silica oversaturated to mildly undersaturated alkalic intrusions consist of gabbro, syenite, nepheline syenite and granite and form massive stocks and arcuate ring complexes.

The term "carbonatite" is presently used to define an igneous rock consisting of 50% or more carbonate (Streckeisen 1980). However, the term is often used in reference to an alkalic complex consisting of a large variety of related silica undersaturated rocks in addition to carbonatite. In Ontario, the strongly silica-undersaturated carbonatite complexes consist dominantly of calcite carbonatite (sövite), silicocarbonatite (carbonate plus oxide, silicate and phosphate minerals) and nepheline-pyroxene rocks (ijolite) and form small circular to elliptical plug-like bodies enveloped in a halo of metasomatic rock (fenites) of varying width (Table 18.1). The kimberlites, described to date in Ontario, are small dike-like bodies which have been outlined by diamond drilling.

The alkalic-carbonatite complexes of Ontario were emplaced during 1 Archean and 3 Proterozoic events. A younger Jurassic (150 million-year-old) event is represented by kimberlite at Kirkland Lake. Numerous lamprophyric diatremes that penetrate the Paleozoic cover rocks of the Hudson Bay Lowland in the area of James Bay (Figure 18.1) may be coeval with the Jurassic kimberlites. All complexes are shown and numbered on Figure 18.1, please refer to Figure 18.1 as the complexes are discussed in the text. Most Proterozoic alkalic-carbonatite intrusions have a close spatial relationship to regional crustal fractures such as the Trans-Superior Tectonic Zone (TSTZ), Kapuskasing Structural Zone (KSZ) and the Ottawa–Bonnechere Graben and its branches (Figure 18.2).

Numerous articles exist dealing with specific topics on individual alkalic or carbonatite intrusions. Regional investigations are restricted to the work of Parsons (1961), Currie (1976), Bell et al. (1982) and Bell and Blenkinsop (1987).

GEOCHRONOLOGY

The geochronology of the alkalic-carbonatite complexes of the Superior Province indicates that the geologic structures hosting the intrusions are long-lived and have been periodically reactivated. The most recent isotopic ages for the complexes, and the method used, are cited in Table 18.2. The oldest alkalic rocks are Archean feldspathoidal syenites and syenites. Most of these Archean complexes are imprecisely dated by K-Ar and Pb-Pb methods, but are approximately 2.7 billion years old. Regional structural control for the Archean Poohbah Lake and Wapikopa River alkalic complexes is not obvious; however, other Archean intrusions such as Sturgeon Narrows may be related to major faults (Sage 1988d, 1988f, 1988n, in prep.). The Archean alkalic complexes may be related to continental cratonization; therefore, they have an origin which is different from the

Table 18.1. Definition of rock and mineral names used in describing carbonatite complexes (*modified from* American Geological Institute 1987; Heinrich 1966; Williams et al. 1954; Streckeisen 1980).

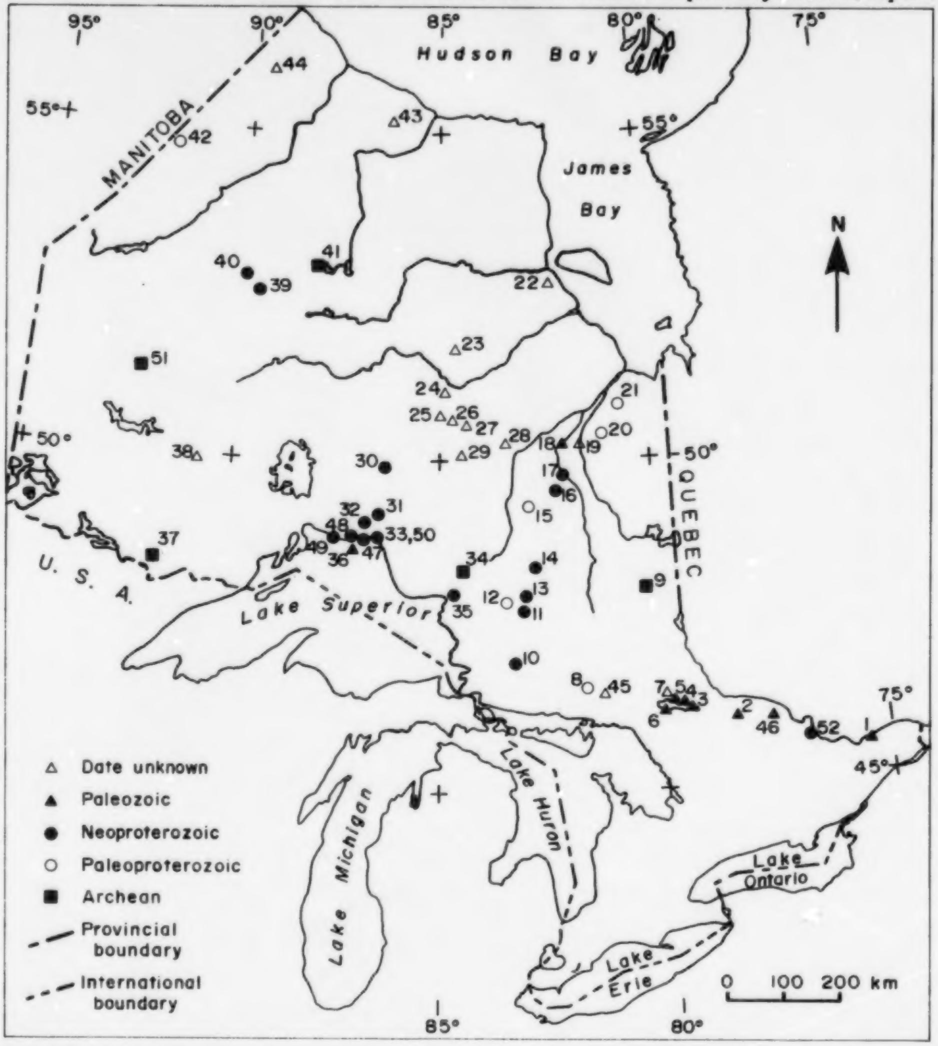
Rock Names	
Alnoite	A lamprophyre chiefly composed of biotite or phlogopite and melilite as essential minerals, commonly with olivine, calcite and clinopyroxene.
Fenite	A rock that has been altered by alkali metasomatism.
Ijolite	A nepheline-pyroxene rock with a nepheline content between 30 and 70%. Rocks containing more than 70% nepheline are classified as urtite and those with less than 30% as melteigite. Some specimens may contain significant amounts of biotite in place of pyroxene. Potassium feldspar content is 10% or less and those rocks with 10% or less nepheline are classified as pyroxenite.
Malignite	A melanocratic nepheline syenite. In general, nepheline, pyroxene and potassium feldspar occur in roughly equal proportions. The potassium feldspar content must exceed 10% or the rock is classified as belonging to the ijolite suite. Both the nepheline and pyroxene content must exceed 10% or the rock would be classified with the syenites. This rock group is transitional between the ijolites and overlaps the syenitic rock groups.
Natrocarbonatite	A rock that is composed of sodium-potassium-calcium carbonates.
Rauhaugite	A rock identical to sövite but with dolomite being the dominant carbonate mineral.
Silicocarbonatite	A carbonatite rock containing 50% or more oxide and silicate minerals. Where the silicate or oxide mineralogy exceeds 90%, various other rock names are applied; i.e., ijolite, biotite, pyroxenite, etc.
Sövite	A carbonatite rock composed of 50% or more calcite. Various mineralogic modifiers are used to classify the sövite, for example, apatite-magnetite sövite, olivine-amphibole sövite, etc.

Mineral Names (Fleisher 1987)	
Akermanite	$Ca_2MgSi_2O_7$, Melilite group
Annite	$KFe_3^{+2}AlSi_3O_{10}(OH,F)_2$, Mica group
Cancrinite	$Na_6Ca_2Al_6Si_6O_{24}(CO_3)_2$, Cancrinite group
Clinohumite	$(Mg,Fe^{+2})_9(SiO_4)_4(F,OH)_2$, Humite group
Edenite	$NaCa_2(Mg,Fe^{+2})_5\,Si_7AlO_{22}(OH)_2$, Amphibole group
Gehlenite	$Ca_2Al(AlSi)O_7$, Melilite group
Hastingsite	$NaCa_2(Fe^{+2}, Mg)_4\,Fe^{+3}Si_6Al_2O_{22}(OH)_2$, Amphibole group
Hydronephelite	A zeolitic alteration of nepheline (Rogers and Kerr 1942)
Kalsilite	$KAlSiO_4$
Melilite	A mineral of the series **Akermanite-Gehlenite**
Pargasite	$NaCa_2(Mg,Fe^{+2})_4\,Al(Si_6Al_2)O_{22}(OH)_2$, Amphibole group
Perovskite	$CaTiO_3$, Perovskite group
Riebeckite	$Na_2(Fe^{+2}, Mg)_3Fe_2^{+3}\,Si_8O_{22}(OH)_2$, Amphibole group
Siderophyllite	$KFe_2^{+2}Al(Al_2Si_2)O_{10}\,(F,OH)_2$, Mica group

younger fault or rift-related intrusions (Sage 1986; Kwon 1986; Bell et al. 1982).

While Archean syenite complexes are not uncommon, Archean feldspathoidal rocks are scarce. The best examples of feldspathoidal syenites in Ontario are the Herman Lake, Sturgeon Narrows and Poohbah Lake complexes, the latter 2 of which have been dated at about 2.7 Ga using Pb-Pb methods (Kwon 1986; Sage 1988d, 1988f, in prep.). Carbonatites of Archean age are unknown in Ontario, with the exception of Springpole Lake (Barron et al. 1989). Barron et al. (1989) described Springpole Lake as an alkalic volcanic complex with minor fluorite-bearing carbonatite, dated by Pb-Pb techniques as Archean in age.

The 1.8 to 1.9 billion-year-old magmatic event is equivalent in age to the Penokean orogenic event (Klasner et al. 1982). Four carbonatite complexes of the 1.8 to 1.9 billion-year-old age group (Borden Township, Cargill Township, Argor and Goldray carbonatite complexes; Sage 1987c, 1988h, 1988k, 1988l) occur within the Kapuskasing Structural Zone. A fifth intrusion, the Carb Lake Carbonatite Complex (Sage 1987f), occurs in northwestern Ontario and may be related to northwest-trending faults. A sixth


95°
90°
85°
80°
75°
Hudson Bay
55°
MANITOBA
James Bay
N
50°
QUEBEC
U. S. A.
Lake Superior
Lake Michigan
Lake Huron
Lake Ontario
Lake Erie
45°
△ Date unknown
▲ Paleozoic
● Neoproterozoic
○ Paleoproterozoic
■ Archean
Provincial boundary
International boundary
0 100 200 km

1. Eastview carbonatite(s)
2. Brent Crater
3. Callander Bay Alkalic Complex
4. Manitou Island Alkalic Complex
5. Burritt Island Alkalic Complex
6. Iron Island Alkalic Complex
7. Lavergne carbonatite
8. Spanish River Carbonatite Complex
9. Otto alkalic complex
10. Seabrook Lake Carbonatite Complex
11. Lackner Lake Alkalic Complex
12. Borden Township Carbonatite Complex
13. Nemegosenda Lake Alkalic Complex
14. Shenango Township alkalic rock
15. Cargill Township Carbonatite Complex
16. Teetzel Township carbonatite
17. Clay–Howells Alkalic Complex
18. Hecla–Kilmer Alkalic Complex
19. Valentine Township Carbonatite Complex
20. Goldray Carbonatite Complex
21. Argor Carbonatite Complex
22. Lawashi River aeromagnetic anomaly
23. Poplar River aeromagnetic anomaly
24. Albany Forks Carbonatite Complex
25. Little Drowning River aeromagnetic anomaly
26. Kingfisher River West aeromagnetic anomaly
27. Kingfisher River East aeromagnetic anomaly
28. Martison Carbonatite Complex
29. Nagagami River Alkalic Complex
30. Chipman Lake fenites and carbonatite dikes
31. Killala Lake Alkalic Complex
32. Prairie Lake Carbonatite Complex
33. Port Coldwell alkalic complex
34. Herman Lake Alkalic Complex
35. Firesand River Carbonatite Complex
36. Slate Islands diatremes
37. Poohbah Lake Complex
38. Sturgeon Narrows Alkalic Complex
39. Schryburt Lake Carbonatite Complex
40. Big Beaver House Carbonatite Complex
41. Wapikopa River Alkalic Complex
42. Carb Lake Carbonatite Complex
43. Gooseberry Brook aeromagnetic anomaly
44. Niskibi Lake aeromagnetic anomaly
45. Nemag and Lusk lakes fenites
46. Allen Lake Carbonatite
47. Dead Horse Creek diatreme
48. McKellar Creek diatreme
49. Gold Range diatreme
50. Neys diatreme
51. Springpole Lake
52. Sullivan Island Carbonatite Complex

Figure 18.1. Location of magmatic events, Superior Province and nearby areas in Ontario—alkalic-carbonatite complexes, aeromagnetically inferred alkalic complexes and fenites.

Table 18.2. Most reliable isotopic age determinations on alkalic and carbonatite complexes of Ontario (*modified from* Sage 1986).

Complex	Age (Ma)	Method	Reference
Archean			
Herman Lake		Geological interpretation	
Poohbah Lake	2706±23	K-Ar	Mitchell 1976
	2672±8	Pb-Pb	Kwon 1986
Sturgeon Narrows	2746±58	K-Ar	Kwon 1986
Wapikopa Lake	2534±147	K-Ar	Sage 1988e
Paleoproterozoic			
Argor	1950	Pb-Pb	Kwon 1986
Borden Township	1894±29	Pb-Pb	Bell et al. 1987
Carb Lake	1826±97	K-Ar	Sage 1988f
Cargill Township	1906±30	U-Pb	Kwon 1986
Goldray	1884±2	U-Pb	Kwon 1986
Spanish River	1838±95	Rb-Sr	Bell and Blenkinsop 1980
McKellar dikes	1653±122	Rb-Sr	Platt and Mitchell 1982
Mesoproterozoic			
Big Beaver House	1109±61	K-Ar	Sage 1987e
Chipman Lake	1029±31	K-Ar	Sage 1985
Clay–Howells	1072±16	Rb-Sr	Bell and Blenkinsop 1980
Dead Horse Creek diatreme		Geologic interpretation	
Firesand River	1008	K-Ar	Gittins et al. 1967
Killala Lake	1098±48	Rb-Sr	Bell and Blenkinsop 1980
Lackner Lake	1092±21	Rb-Sr	Bell and Blenkinsop 1980
Nemegosenda Lake	1015±63	Rb-Sr	Bell and Blenkinsop 1980
Port Coldwell	1108±1	U-Pb	Heaman and Machado 1987
Prairie Lake	1165±30	Rb-Sr	Pollock 1987
Schryburt Lake	1145±74	K-Ar	Sage 1988i
Seabrook Lake	1092–1109	K-Ar	Gittins et al. 1967
Shenango Township	1047±15	Rb-Sr	Bell et al. 1979 Bell and Blenkinsop 1980
Teetzel Township	1155	K-Ar	Wanless et al. 1973
Valentine Township		Geologic interpretation	
Cambrian			
Brent	414±20	K-Ar	Hartung et al. 1971
Hecla–Kilmer	450	Rb-Sr	Bell and Blenkinsop 1980

Table 18.2. Most reliable isotopic age determinations on alkalic and carbonatite complexes of Ontario (*modified from* Sage 1986).

Complex	Age (Ma)	Method	Reference
Manitou Island	560	K-Ar	Gittins et al. 1967
Newman Island	570	K-Ar	Gittins et al. 1967
Jurassic			
Upper Canada kimberlite	158±2	K-Ar	Heaman 1989
Mafic alkalic diatremes	150±8 180±9	K-Ar	Janse et al. 1986
Miscellaneous			
Slate Islands	282±11 310±18	K-Ar	Sage 1983

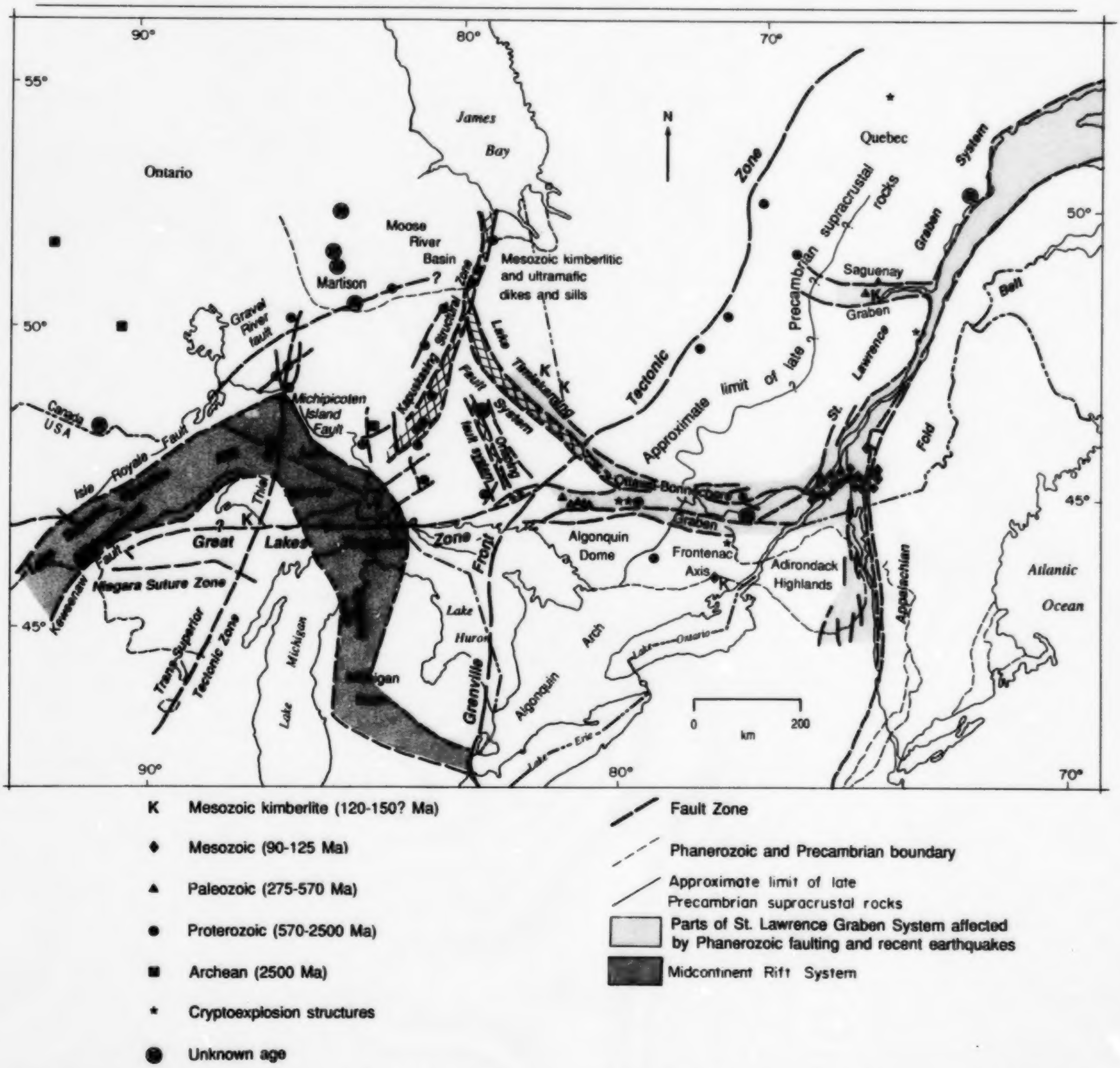

90°
80°
70°
55°
50°
45°
James
Bay
Ontario
Quebec
N
Moose
River
Basin
Martison
Mesozoic kimberlitic
and ultramafic
dikes and sills
Gravel
River
fault
Canada
USA
Isle Royale Fault
Michipicoten
Island
Fault
Kapuskasing Structural Zone
Lake Timiskaming Fault System
Ottawa-Bonnechere Graben
Great Lakes Zone
Thiel
Keweenaw Fault
Niagara Suture Zone
Trans-Superior Tectonic Zone
Lake Michigan
Michigan
Lake Huron
Grenville Front
Algonquin Arch
Algonquin
Dome
Frontenac
Axis
Adirondack
Highlands
Lake Ontario
Lake Erie
Tectonic Zone
Approximate limit of late Precambrian supracrustal rocks
Saguenay
Graben
St. Lawrence Graben System
Appalachian Fold Belt
Atlantic
Ocean
0
km
200
K Mesozoic kimberlite (120-150? Ma)
◆ Mesozoic (90-125 Ma)
▲ Paleozoic (275-570 Ma)
● Proterozoic (570-2500 Ma)
■ Archean (2500 Ma)
* Cryptoexplosion structures
● Unknown age
Fault Zone
Phanerozoic and Precambrian boundary
Approximate limit of late
Precambrian supracrustal rocks
Parts of St. Lawrence Graben System affected
by Phanerozoic faulting and recent earthquakes
Midcontinent Rift System

Figure 18.2. Major regional structures and alkalic intrusions within eastern Canada (all alkalic rock and carbonatite intrusions are not shown) (*modified from* Lumbers 1978 and Brummer 1978).

intrusion, the Spanish River Carbonatite Complex (Sage 1987d), occurs northwest of the Sudbury Basin and is related to faults external to the basin.

Using Rb-Sr techniques, Bell and Blenkinsop (1980) dated the Spanish River Carbonatite Complex at 1838±95 Ma which is within error of the age of the Sudbury Igneous Complex dated at 1850±1 Ma by the U-Pb technique (Krogh et al. 1984). Existing U-Pb ages suggest that the 1.8 to 1.9 billion-year-old carbonatite intrusive event was of short duration (*see* Table 18.2).

A major, and, perhaps the dominant alkalic-magmatic event took place along major faults in the Superior Province between 1.0 and 1.2 Ga (*see* Table 18.2) and includes both alkalic and carbonatite complexes. Intrusions of this age group are commonly emplaced in proximity to older alkalic intrusions. For example, the 1.0 billion-year-old Nemegosenda Lake Alkalic Complex is emplaced near the 1.9 billion-year-old Borden Township Carbonatite Complex (Sage 1987b, 1987c). The 1.0 to 1.2 billion- year- old event is equivalent in age to the Keweenawan rift event and to the Grenville deformation (Klasner et al. 1982).

An early Cambrian (0.5 to 0.6 billion-year-old) alkalic and carbonatite event occurred in the area of Lake Nipissing and is related to continental rifting (Doig 1970; Kumarapeli 1978; *see* Figure 18.2). The Hecla–Kilmer Alkalic Complex, at the north end of the Kapuskasing Structural Zone, is also early Cambrian (Sage 1988m).

The most recent alkalic events are the following: alkalic ultramafic diatreme emplacement in the Hudson Bay Lowlands, kimberlite diking on the extrapolated extension of the Lake Timiskaming Fault System (*see* Figure 18.2) at 150 to 180 Ma (Jurassic) (Janse et al. 1986) and 120 million-year-old lamprophyre dikes in the St. Lawrence Platform (Barnett et al. 1984). The St. Lawrence Platform occurrences are further described in the Phanerozoic chapter of this volume. This event is related to reactivation of earlier rift structures (Kumarapeli 1978).

REGIONAL STRUCTURE

Structural geology provides the explanation for the areal distribution of most alkalic intrusions. Within Ontario, there are 3 major structural zones controlling post-Archean alkalic magmatism. These structural zones are the Trans-Superior Tectonic Zone, the Kapuskasing Structural Zone and the Ottawa–Bonnechere Graben.

Trans-Superior Tectonic Zone

The Trans-Superior Tectonic Zone (TSTZ) is a north-northeast-trending fault zone which extends for a distance greater than 600 km and includes the Thiel fault in Lake Superior (Klasner et al. 1982). The TSTZ has also been referred to as the Big Bay–Ashburton Bay Fault by Sage (1978). The Thiel fault part of the TSTZ, which has been outlined seismically in Lake Superior (Smith et al. 1966; Cannon et al. 1989), lies along a topographic high (Hough 1958) and is marked by a disruption in aeromagnetic (Hinze et al. 1966) and gravity (Klasner et al. 1979) contours.

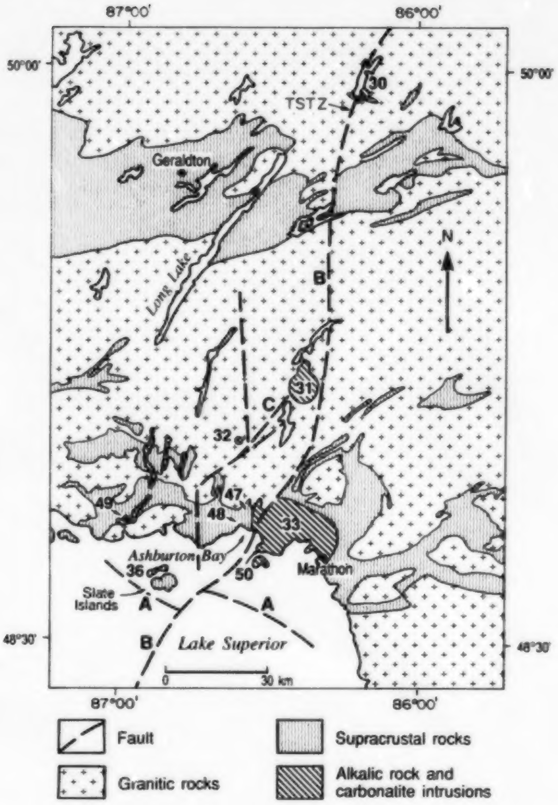


Figure 18.3. Schematic sketch map indicating the location of alkalic rock and carbonatite intrusions along the northern part of the Trans–Superior Tectonic Zone (TSTZ). The major carbonatite and alkalic rock intrusions and major regional faults are: 30-Chipman Lake fenites and carbonatite dikes; 31-Killala Lake Alkalic Complex; 32-Prairie Lake Carbonatite Complex; 33-Port Coldwell alkalic complex; 36-Slate Islands diatremes and carbonatite dike; 47-Dead Horse Creek diatreme; 48-McKellar Creek diatreme; 49-Gold Range diatreme; 50-Neys diatreme; A-Michipicoten Island Fault; B-Thiel fault and its extrapolated northern extension; C-Killala Lake deformation zone.

The TSTZ intersects the geophysically defined Michipicoten Island Fault immediately south of the Slate Islands (Halls and West 1971; Figure 18.3). From the Slate Islands, a topographic lineament can be traced northward to the Prairie Lake Carbonatite Complex, while the main trend of the TSTZ is mapped farther to the east toward the Port Coldwell alkalic complex (*see* Figure 18.3). From the Port Coldwell alkalic complex, prominent topographic lineaments can be traced north to the Killala Lake Alkalic Complex and the Chipman Lake fenites and carbonatite dikes (Sage 1985; *see* Figure 18.3). Near the Killala Lake Alkalic Complex, the prominent north-striking topographic lineament has been interpreted as a fault (Coates 1970). Sage (1985) recognized 0.8 km left-lateral faulting along the lineament at Chipman Lake.

Williams (1989) has recently identified a zone of deformation named the Killala Lake deformation zone between the Prairie Lake Carbonatite and Killala Lake alkalic complexes. The deformation zone, which is several kilometres wide, defines the boundary between the Wawa and Quetico subprovinces and displays tectonic indicators showing a dextral horizontal component of movement (Williams 1989).

The major alkalic event along the TSTZ occurred from 1.0 to 1.2 Ga. In addition, Sage (1983) identified lamprophyre dikes dated at approximately 300 Ma on the Slate Islands by the K-Ar method, and Platt and Mitchell (1982) reported 1653±122 million-year-old lamprophyres dated by the Rb-Sr method in the same region. Work by Heaman and Krogh (1986) indicate that at least some of the lamprophyre dikes in this area interpreted to be of 1600 million year age are younger and have been dated at 1130 Ma by the U-Pb method; therefore, they are similar in age to the main alkalic event in the region.

Kapuskasing Structural Zone

The Kapuskasing Structural Zone (KSZ) extends from the east shore of Lake Superior northeast to James Bay (Figure 18.4). The KSZ is poorly defined along the east shore of Lake Superior but becomes better defined towards James Bay. The KSZ crosscuts an east-trending fabric within Archean rocks of the Superior Province and is subparallel to the TSTZ (*see* Figure 18.2).

The KSZ is characterized by a northeast-striking linear aeromagnetic pattern (400 to 600 gammas above regional background) and positive gravity highs (up to 20 mgal) (Innis 1960; ODM–GSC 1970; GSC 1984). Numerous alkalic and carbonatite intrusions occur along this structure (*see* Figure 18.4).

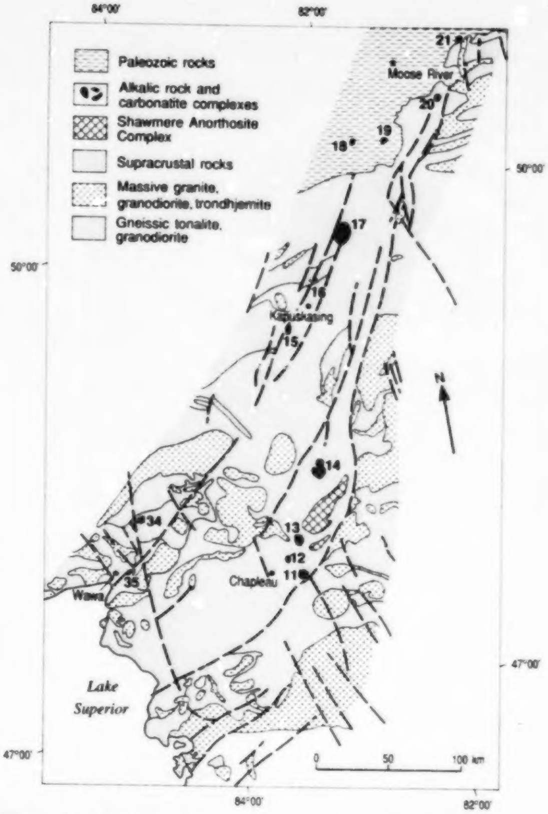


Figure 18.4. Schematic sketch of Kapuskasing Structural Zone (KSZ) and location of alkalic rock and carbonatite intrusions (geology *modified from* Card 1982). 11-Lackner Lake Alkalic Complex; 12-Borden Township Carbonatite Complex; 13-Nemegosenda Lake Alkalic Complex; 14-Shenango Township alkalic rock; 15-Cargill Township Carbonatite Complex; 16-Teetzel Township carbonatite; 17-Clay–Howells Alkalic Complex; 18-Hecla–Kilmer Alkalic Complex; 19-Valentine Township Carbonatite Complex; 20-Goldray Carbonatite Complex; 21-Argor Carbonatite Complex; 34-Herman Lake Alkalic Complex; 35-Firesand River Carbonatite Complex.

The KSZ has been interpreted as an upwarp of the Conrad Discontinuity (Wilson and Brisbin 1965; Bennett et al. 1967; Thurston et al. 1977), a product of collision of the Churchill and Superior cratons in Paleoproterozoic time (Gibb 1978) and as a deep transcurrent shear (Watson 1980). More recently, Percival and Card (1983) have proposed that the KSZ is an east-verging thrust fault which has exposed an oblique section through 20 km of uplifted Archean crust. Granulite-facies rocks of the KSZ are juxtaposed against greenschist-facies rocks of the Abitibi Subprovince along the Ivanhoe Lake cataclastic zone. The KSZ is characterized by a high-grade gneiss terrain and grades westward into a central gneiss terrain and then into low-grade terrain of east-west-striking linear belts composed of supracrustal rocks.

In addition to the major fault which forms the east boundary of the KSZ, 3 major northeast-striking faults dip 60° to 70° northwest and are present within the uplift (Percival and McGrath 1986). These internal faults are west-side-down with displacement of 7 to 10 km and result from a late tensional event that followed the compressional uplift (Percival and McGrath 1986; Percival 1987).

Northey and West (1986) have interpreted seismic refraction data to indicate a crustal thickness of 48 km below the KSZ, which thins slightly to the west and the east. The interpretation of seismic data is consistent with the upthrust of mid-crustal rocks along a west-dipping listric fault (Northey and West 1986; Boland et al. 1988).

The precise age of the KSZ is uncertain (Watson 1980; Percival and Card 1983), but the 1800 to 1900 million-year-old carbonatite intrusions are exposed at a high structural level and are enclosed within high-grade deep level gneisses. The KSZ has therefore formed prior to the 1800 to 1900 million-year-old carbonatite intrusion event; however, the KSZ remains seismically active (Forsyth and Morel 1982).

Alkaline magmatism along the KSZ took place during the Penokean and Grenville-Keweenawan orogenic events and in the Neoproterozoic to Early Cambrian and Jurassic periods. Woolley (1989) has documented the worldwide spatial and temporal distribution of carbonatite with doming, linear structures and orogenesis.

Ottawa–Bonnechere Graben

The Ottawa–Bonnechere Graben, a branch of the St. Lawrence Graben System, lies along the Ontario–Quebec border and is approximately 60 km wide (Kumarapeli and Saull 1966; *see* Figure 18.2). The fault system splits at Lake Nipissing into a west-trending branch and the northwest-trending Lake Timiskaming Fault System (Forsyth 1981; Lovell and Caine 1970; Figure 18.5). Along the graben system, lamprophyre dikes are common. Within the region of the bifurcation of the rift, 7 alkalic-carbonatite complexes are present (*see* Figure 18.5). At the southern end of the Ottawa–Bonnechere Graben, in the region where it joins with the St. Lawrence Graben System, numerous small carbonatite intrusions are present, but these occur mostly on the Quebec side of the Ontario–Quebec border (Hogarth and Rushforth 1986).

Kumarapeli (1978) suggested that the St. Lawrence Graben System was related to the opening of the Atlantic Ocean, and Doig (1970) correlated the rifting in eastern Canada with rifting in Greenland and northern Europe. Doming and arching have not been found to be associated with the Ottawa–Bonnechere Graben (Doig 1970), unlike more classical African rift models (Bailey 1964, 1974).

The Ottawa–Bonnechere Graben is at present seismically active (Poole et al. 1970; Kumarapeli 1987; Adams 1989), but in the vicinity of Ottawa, the main zone of seismicity is oblique to the graben, and this oblique orientation remains largely unexplained (Forsyth 1981; Kumarapeli 1987; Adams 1989). Adams (1989) correlated the zone of earthquake activity along the Ottawa Valley to rifting and suggests that the oblique zone of earthquake activity may be related to a migrating hot spot. Seismic reflection and refraction surveys across the Ottawa–Bonnechere Graben indicate that the Moho is poorly defined and the graben may have upper mantle material emplaced within it (Mereu et al. 1986).

LOCAL STRUCTURE

Alkalic Complexes

The alkalic complexes are emplaced at inflections in the trends of major structural zones or at sites of cross faulting, both of which provide dilatant zones permitting emplacement of alkaline magmas from the lower crust and upper mantle. The ring complexes, such as the Port Coldwell and Killala Lake alkalic complexes, have formed by ring fracturing and caldera subsidence. The Port Coldwell alkalic complex consists of 3 superimposed ring complexes younging southward along the TSTZ (Currie 1980; Sage 1986). The Killala Lake Alkalic Complex is composed of a single ring (Sage 1988e). Relatively homogeneous stocks lacking arcuate distribution of rock types, such as the Clay–Howells Alkalic Complex, may represent the expanded upper part of dike-like intrusions. Within alkalic ring complexes, early mafic units commonly occupy ring fractures (Figures 18.6 and 18.7). Flow banding, mafic segregation bands and trachyoidal textures conform to the outline of the arcuate rock units found in ring complexes.

The abundance of xenolithic blocks, roof pendants and wall-rock septa on the north side of the Port Coldwell and Killala Lake alkalic complexes, suggests that these complexes are exposed at high structural levels on their northern flanks. The absence of similar structures on the south sides of these complexes suggests a deeper structural level of exposure and that they may plunge steeply south along the TSTZ (Sage 1986, 1988e).

Carbonatites and Kimberlites

Carbonatites may occur at inflections or intersections of 2 structural trends within the regional structures. Since carbonatites represent mantle-derived melts (Bell and Blenkinsop 1987), these regional structures are likely rooted in the lower crust-upper mantle and provide a means by which carbonatites reach high crustal levels. Carbonatites rarely exceed 3 to 4 km in diameter and commonly display a heterogeneous mixture of rock types. Internal structures such as cumulate banding, flow banding or mineralogic layering are subvertical and crudely outline the complex.

Kimberlites and compositionally similar units form dikes associated with deep crustal fractures such as the Lake Timiskaming Fault System which also contain alkalic and carbonatite complexes. Kimberlitic rocks in diatreme structures have been identified from the Hudson Bay Lowland in the area of James Bay, but structural control for their emplacement has not been defined (Janse et al. 1986).

ISOTOPE GEOLOGY

A study of Rb-Sr isotopic systems in Ontario carbonatites by Bell et al. (1982) indicated a common depleted mantle source region that has behaved as a closed system for at least 2.6 billion years. Subsequently, Bell and Blenkinsop (1987) postulated a mantle extraction age of 2.9 Ga by studying neodymium and strontium initial isotopic ratios in carbonatites. Nelson et al. (1988) studied the isotopes of several carbonatites outside Ontario and concluded that the carbonatites were likely to be petrogenetically related to a basaltic kimberlite magma. The carbonatites were generated from a CO_2-rich and volatile-rich oceanic lithosphere which was repeatedly replenished in incompatible elements. The model of Nelson et al. (1988) is one of an open system as opposed to the closed system proposed by Bell and Blenkinsop (1987).

Strontium, neodymium and lead isotopic studies of the 1.9 billion-year-old Borden Township, Cargill Township, and Spanish River carbonatite complexes suggest derivation from a LIL-depleted mantle with little or no crustal contamination (Bell et al. 1982, 1987). The elevated LIL abundances in the carbonatites in conjunction with the depleted isotopic ratios suggest that the mantle source was enriched or metasomatized in LIL shortly before melting. The Borden Township Carbonatite Complex (1.9 Ga) lies 13 km from the Nemegosenda Lake (1.0 billion-year-old) and Lackner Lake (1.0 billion-year-old) alkalic complexes.

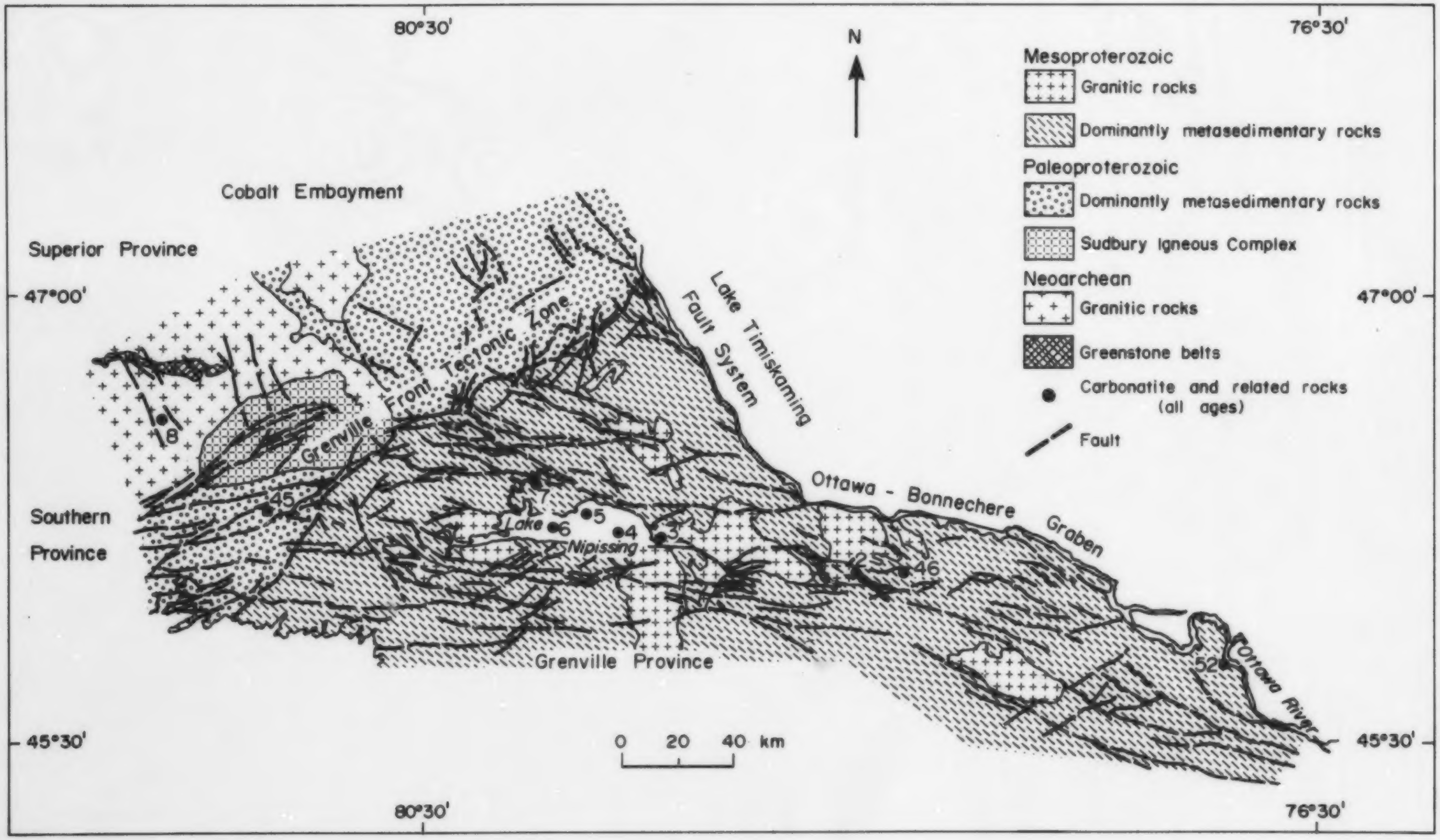

80°30'
76°30'
N
Mesoproterozoic
Granitic rocks
Dominantly metasedimentary rocks
Paleoproterozoic
Dominantly metasedimentary rocks
Sudbury Igneous Complex
Neoarchean
Granitic rocks
Greenstone belts
Carbonatite and related rocks (all ages)
Fault
Cobalt Embayment
Superior Province
47°00'
Lake Timiskaming Fault System
Grenville Front Tectonic Zone
8
45
Southern Province
7
5
Lake 6
4
Nipissing
3
2
46
Ottawa - Bonnechere Graben
Grenville Province
52
Ottawa River
45°30'
0 20 40 km

Figure 18.5. Schematic diagram showing the location of alkalic rock and carbonatite complexes at the northern end of the Ottawa–Bonnechere Graben (*modified from* Ayres et al. 1970; Lumbers and Milne 1979). 2-Brent Crater; 3-Callandar Bay Alkalic Complex; 4-Manitou Island Alkalic Complex; 5-Burritt Island Alkalic Complex; 6-Iron Island Alkalic Complex; 7-Lavergne carbonatite; 8-Spanish River Carbonatite Complex; 45-Nemag and Lusk lakes fenites; 46-Allen Lake Carbonatite; 52-Sullivan Island Carbonatite Complex.

The Borden Township Carbonatite Complex and Lackner Lake Alkalic Complex were both derived from depleted mantle sources (Bell et al. 1982, 1987), whereas the Nemegosenda Lake Alkalic Complex was not (Bell et al. 1982). This close spatial association of alkalic-carbonatite intrusions of greatly differing age suggests that metasomatism replenished the fusible components, or that different regions of the mantle were involved in magma genesis.

GEOLOGY

Geology of the alkalic rocks is discussed in 3 groups: alkalic rocks, carbonatites and kimberlites. Two Ontario alkalic complexes, the Lackner Lake and Nemegosenda Lake alkalic complexes, are ring complexes that contain rock types found in both carbonatite and alkalic complexes and may represent magmas intermediate in composition between those forming alkalic complexes and those forming carbonatite complexes (Sage 1987b, 1988c). Table 18.1 presents definitions of selected rock and mineral types which may not be familiar to the reader.

Rb-Sr and Nd-Sm isotope data indicate that the alkalic and carbonatite complexes of Ontario are of lower crust–upper mantle origin (Powell and Bell 1974; Bell et al. 1979, 1982, 1987; Bell and Blenkinsop 1980, 1987). Crystal fractionation of these alkaline magmas has been identified at the Port Coldwell alkalic complex by Whittaker (1976), Mitchell and Platt (1978, 1982) and at the Argor and Cargill Township carbonatite complexes by Twyman (1983). At the Argor Carbonatite Complex, Twyman (1983) considered crystal fractionation as the cause for its high-niobium content.

The generation of alkaline and carbonatite magmas may result from low degrees of partial melting of the mantle or alternatively by higher degrees of partial melting of metasomatized mantle (Boettcher and O'Neil 1980; Nelson et al. 1988). Strontium and neodymium isotopic studies by Bell and Blenkinsop (1987) that indicate depleted mantle sources, suggest mantle metasomatism must immediately precede melting. The presence of several closely spaced alkalic and carbonatite complexes of widely varying age implies that metasomatism replenished the mantle volatile components that were removed during preceding magmatic episodes (Bailey 1982; Erlank et al. 1982). Mantle metasomatism is presently perceived as an important process in generating alkaline magmas (Haggerty 1989; Meen et al. 1989).

Alkalic Complexes

Archean alkalic complexes such as the malignite unit (*see* Table 18.1) at the Poohbah Lake Complex (Sage 1988d) display an arcuate lithologic distribution; however, other complexes such as Sturgeon Narrows do not show arcuate units. The greatest lithologic variety occurs within Proterozoic alkalic ring complexes; the best example of which is the Port Coldwell alkalic complex closely followed by the Killala Lake Alkalic Complex (Puskas 1967; Currie 1980; Sage 1986, 1988e; Figures 18.8 and 18.9). These complexes range in composition from olivine gabbro to granite.

The oldest lithologic units within Proterozoic alkalic complexes of Ontario are the olivine gabbros to diorites. These gabbros, which occur around the margins of the complexes, commonly display well-developed mineralogical

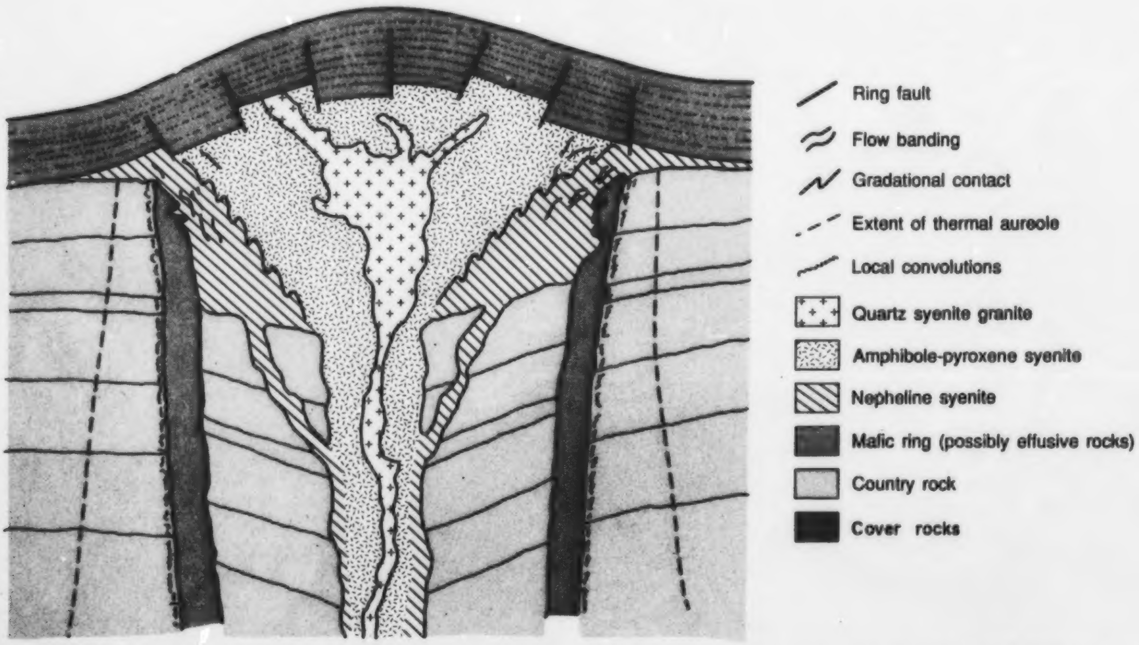


Figure 18.6. Idealized vertical cross section through a ring complex. The model is based on regional mapping of the Port Coldwell and Killala Lake alkalic complexes (Sage 1986).

banding and cumulus textures. Feldspathoidal syenites, which intrude the gabbros, are the next youngest unit in alkalic complexes. Nepheline is the primary feldspathoid and occurs interstitially to potassium feldspar and sodium-iron amphibole. The rocks are medium grained, but local pegmatitic phases are present. The nepheline may be altered to hydronephelite or cancrinite (*see* Table 18.1).

Amphibole-pyroxene and pyroxene syenites are common within alkalic complexes, are dominant at the present level of exposure in both Archean and Proterozoic intrusions and follow the feldspathoidal syenites in order of emplacement. The syenites are medium to coarse grained and locally have well-developed mafic banding such as in the Port Coldwell and Killala Lake alkalic complexes. Within the Proterozoic intrusions, quartz is locally abundant and the syenites are gradational into quartz syenite with amphibole as the principal mafic mineral.

PORT COLDWELL

The best known and most studied alkalic intrusion is the Port Coldwell alkalic complex located on the northeast shore of Lake Superior (*see* Figure 18.8). The Port Coldwell alkalic complex is dated at 1108±1 Ma (Heaman and Machado 1987) and intrudes Archean rocks of the Abitibi and Wawa subprovinces.

Currie (1980) recognized that the complex is not a single ring, but that it represents the superposition of 3 ring complexes. The 3 ring complexes become progressively younger in a southwest direction along the TSTZ. The oldest, the eastern ring complex, is centred near Craddock Cove in Peninsula Bay and is composed of an outer ring of gabbro and an internal sheet-like cone of pyroxene syenite. The second oldest ring complex, the western ring complex, is centred approximately 4 km north of the community of Coldwell and consists of an outer ring of alkalic gabbro to diorite and a core of nepheline syenite, pyroxene syenite, amphibole syenite and quartz syenite. The younger, southern ring complex, is centred on Pic Island. The Pic Island intrusive centre is represented by arcuate sheets of nepheline syenite, amphibole syenite and quartz syenite. Fine-grained syenite and various lamprophyre dikes are common. Based on gravity and aeromagnetic data, Currie (1980) and Mitchell et al. (1983) proposed that the syenitic rocks overlie rocks that are much more mafic in composition and that the complex may have its roots in the lower crust and upper mantle.

Currie (1980) has postulated a basaltic parent for the Port Coldwell alkalic complex, and Mitchell and Platt

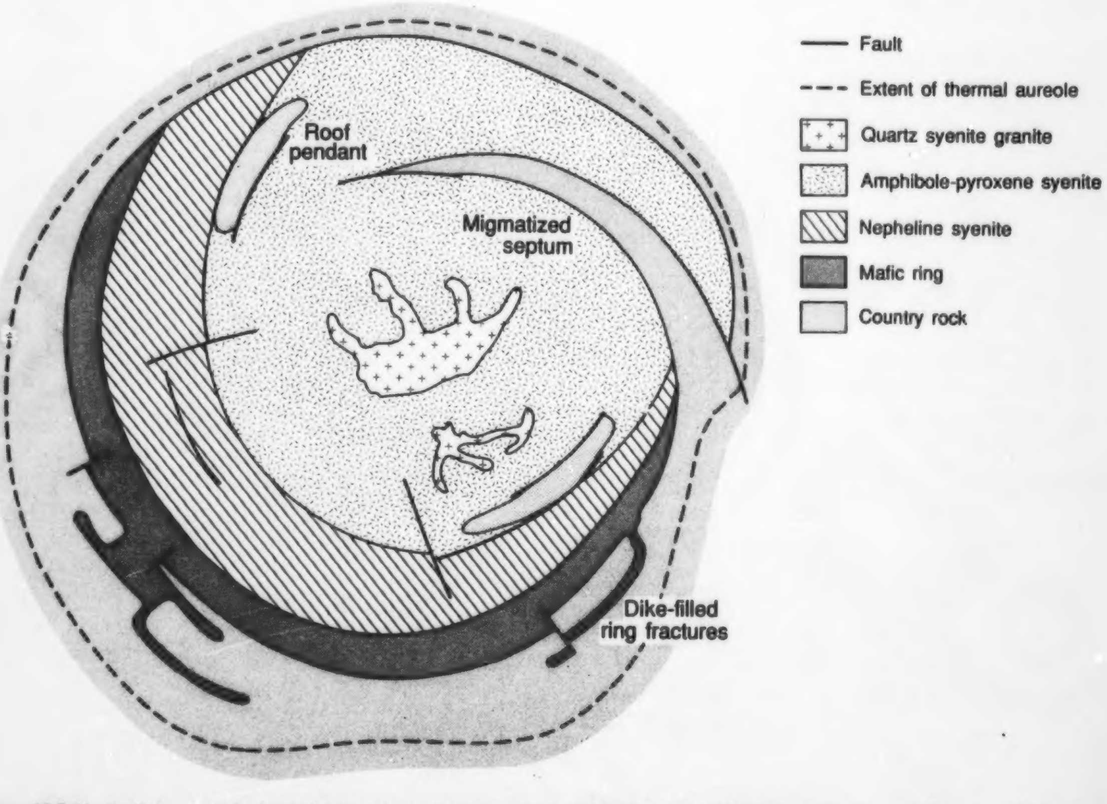


Figure 18.7. Idealized plan view of a simple ring complex based on mapping at the Port Coldwell and Killala Lake alkalic complexes (Sage 1986). Two ages of faults cut the complex. Older ring faults are most common along and parallel to rock unit contacts, while younger faults cut the complex at roughly 90° to the ring faults.

(1982) proposed an alkali basaltic magma as parental to the nepheline syenite phases of the Port Coldwell alkalic complex.

The rocks described as hornfels by Puskas (1967) and lamprophyric gabbro by Currie (1980) occur along the north margin of the complex and near Craddock Cove on Lake Superior. These rocks represent down-dropped blocks of roof material along ring fractures. The amoeboid amygdaloidal filling and relict primary textures suggest that some of the hornfels may represent former Mesoproterozoic flows that might have formed a cover to the intruding alkaline magma. These flows may be the extrusive equivalent of the outer gabbro ring.

The Neys diatreme, located in Neys Peninsula, has been interpreted as a high-level breccia pipe analogous to those found in the porphyry copper environment (Sage 1982). This diatreme contains rounded to angular blocks of various rock types that form the Port Coldwell alkalic complex. The diatreme is likely formed by rapid release of volatiles in a cupola region of the intruding syenite magma. West of the complex, the McKellar Creek and Dead Horse Creek diatremes display potassium, iron, REEs, phosphorus, niobium, uranium and thorium metasomatism which may be analogous to distal fenitization by a carbonatite magma (Sage 1982). The diatreme structures west of the Port Coldwell alkalic complex may not be directly related to the complex, since rare clasts probably derived from the Port Coldwell alkalic complex, are present in the Dead Horse Creek diatreme.

Phases of coarse-grained syenite and its pegmatitic equivalents contain miarolitic cavities lined with crystals of quartz and feldspar which imply a high-volatile content or a low-confining pressure.

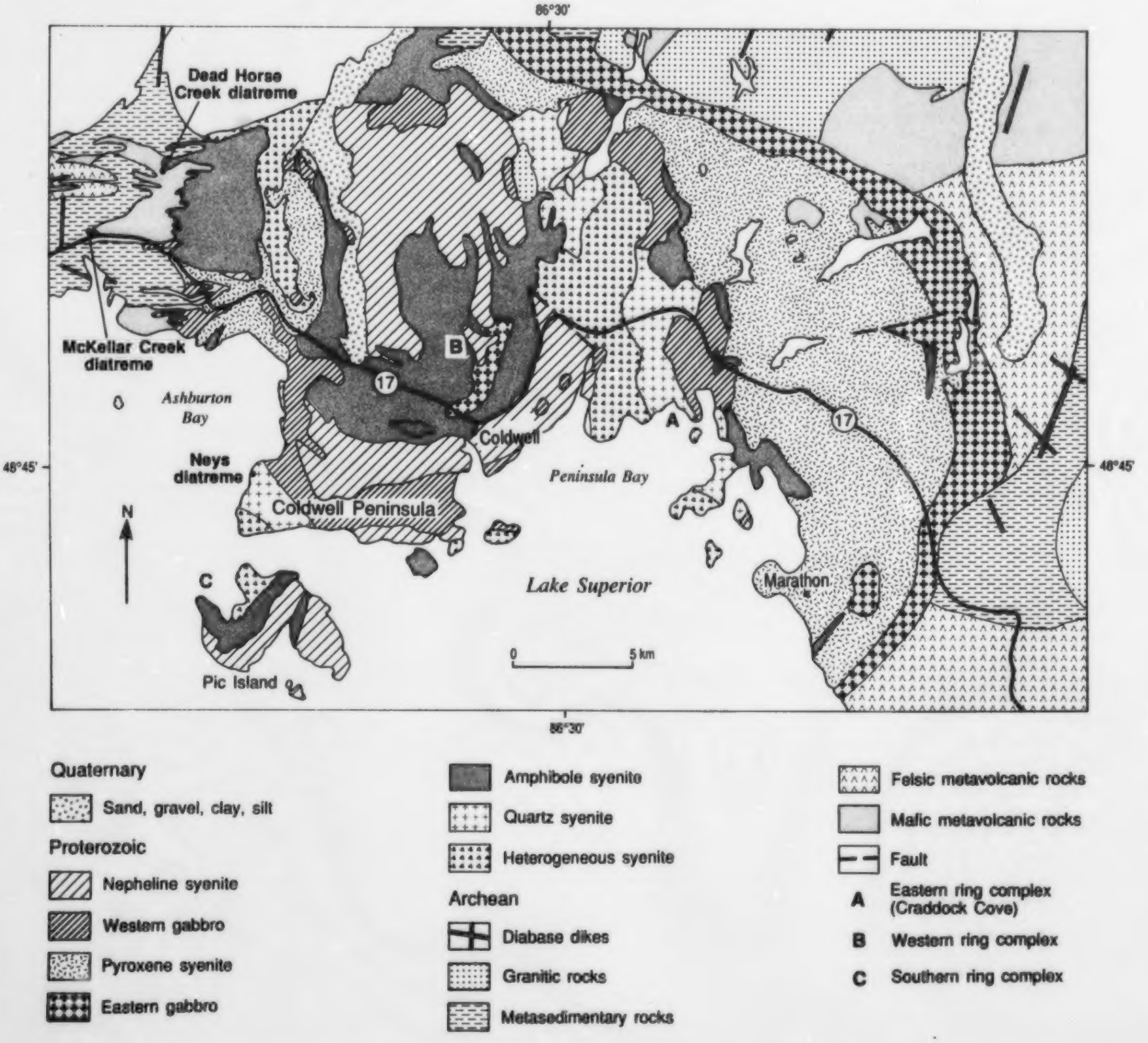

86°30'
Dead Horse Creek diatreme
McKellar Creek diatreme
Ashburton Bay
B
17
Coldwell
A
17
48°45'
Neys diatreme
Coldwell Peninsula
Peninsula Bay
48°45'
N
C
Lake Superior
Marathon
0
5 km
Pic Island
86°30'
Quaternary
Sand, gravel, clay, silt
Proterozoic
Nepheline syenite
Western gabbro
Pyroxene syenite
Eastern gabbro
Amphibole syenite
Quartz syenite
Heterogeneous syenite
Archean
Diabase dikes
Granitic rocks
Metasedimentary rocks
Felsic metavolcanic rocks
Mafic metavolcanic rocks
Fault
A Eastern ring complex (Craddock Cove)
B Western ring complex
C Southern ring complex

Figure 18.8. Schematic sketch of the geology of the Port Coldwell alkalic complex, *modified from* Currie (1980).

Wilkinson (1983) determined that for gabbros within the complex, the compositional range of olivine is Fo_{36-67} and clinopyroxene is Wo_{43-46}, En_{32-43}, Fs_{13-21}. Olivines on the north rim of the complex show a limited compositional range of Fo_{32-36} and on the basis of microprobe data, Whittaker (1976), classified the clinopyroxene as ferroaugite and more magnesium-rich than that found in the associated pyroxene syenites. Microprobe analysis of the plagioclase indicates a compositional range of labradorite to andesine (Whittaker 1976).

Mitchell and Platt (1982) have determined that in the nepheline syenite, the clinopyroxene has a composition of $Ac_{6.8-7.9}Hd_{13.1-62.6}Di_{7.1-71.4}$ and the amphibole, a composition from magnesian hastingsite (hornblende) to hastingsite, to hastingsite (hornblende) to ferro-edenite (hornblende) (*see* Table 18.1; Mitchell and Platt 1982). Associated biotite varies from titaniferous biotite to siderophyllite-annite and the nepheline contains 16.0 to 18.9% kalsilite (Mitchell and Platt 1982; *see* Table 18.1). The alkali feldspars vary from $Or_{28}Ab_{68}An_4$ to $Or_{82}Ab_{18}$ (Mitchell and Platt 1982).

Mitchell and Platt (1982) suggested that the rocks did not undergo long-term subsolidus re-equilibrium as high-temperature mineralogy is preserved and they further proposed that the syenites are likely to be the product of extensive fractional crystallization of alkali basalt magma.

Whittaker (1976) found that the minor olivine in the pyroxene syenite at the Port Coldwell alkalic complex has a compositional range of Fo_{2-6} and that the clinopyroxenes are ferroaugite with a much lower magnesium content than the gabbros. With increasing fractionation, minor secondary amphibole ranges from edentic to ferrorichterite in the exposed syenite section of syenite (Whittaker 1976). Secondary biotite in the syenites is more iron-rich than that in the gabbros. Secondary biotite increases in the annite component with increasing fractionation of the pyroxene syenite (Whittaker 1976).

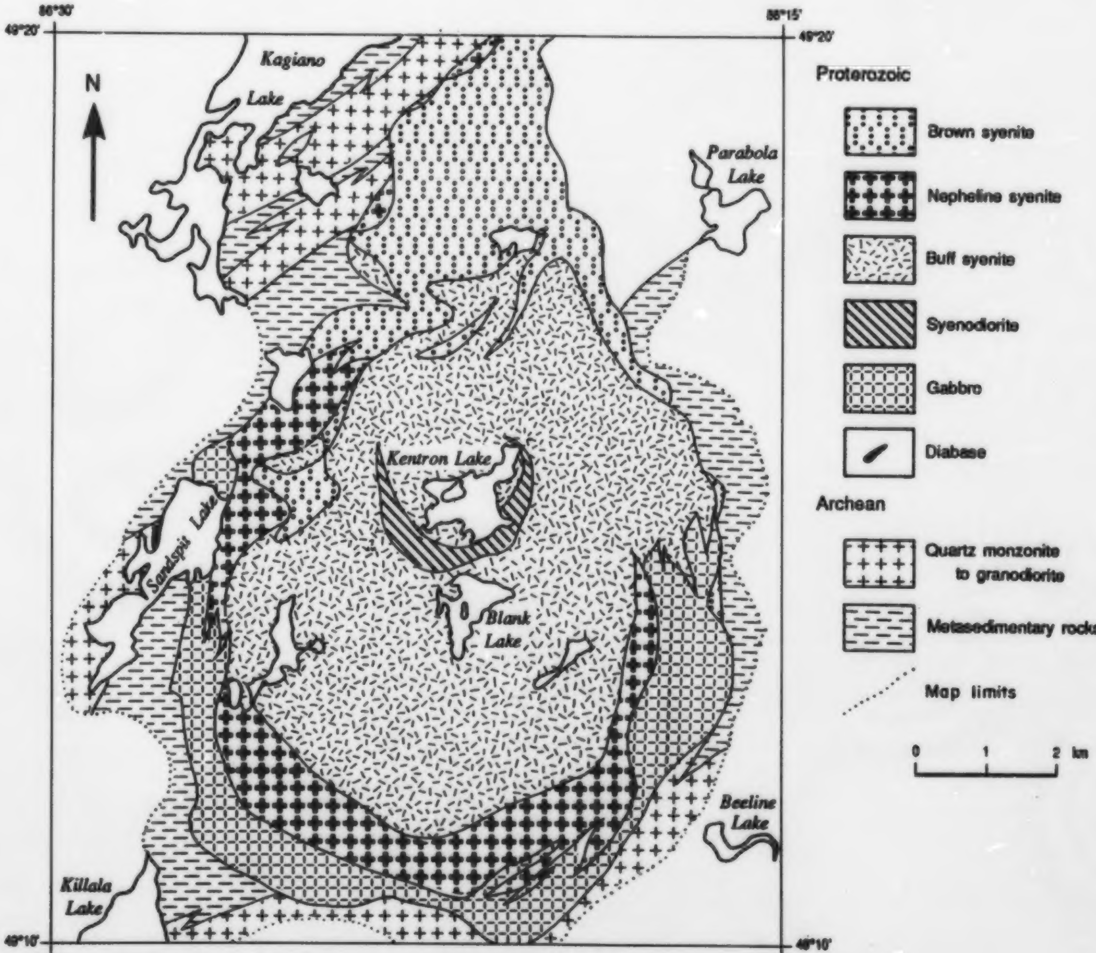

86°30'
86°15'
49°20'
49°10'
N
Kagiano Lake
Parabola Lake
Kentron Lake
Sandspit Lake
Blank Lake
Beeline Lake
Killala Lake
Proterozoic
Brown syenite
Nepheline syenite
Buff syenite
Syenodiorite
Gabbro
Diabase
Archean
Quartz monzonite to granodiorite
Metasedimentary rocks
Map limits
0
1
2 km

Figure 18.9. Schematic sketch map of the geology of the Killala Lake Alkalic Complex (Sage 1988e).

Mitchell and Platt (1978) found the olivine to vary from Fo_{17} to Fo_7 and the pyroxene to have a composition of $Di_{50}Hd_{45}Ac_5$ to $Di_5Hd_{90}Ac_5$ in the least fractionated syenite. More evolved syenites had olivine compositions of Fo_7 and clinopyroxene compositions of $Di_5Hd_{80}Ac_{15}$ to $Ac_{50}Hd_{50}$ (Mitchell and Platt 1978). On the basis of probe chemistry, the amphiboles had a composition of ferrorichteritic katophorite (sodium- to calcium-rich amphibole) to ferrorichterite (sodium- to iron-rich amphibole) (Mitchell and Platt 1978). Mineral compositions and rock textures were used by Mitchell and Platt (1978) to suggest a model of open-system fractionation of an evolved magma leading to late-stage iron-rich, oversaturated, peralkaline liquids.

The plagioclase within the pyroxene syenite is An_{27-37} and decreases in relative abundance up-section with a concomitant increase in microperthite (Whittaker 1976). The youngest phases within the alkalic complexes are the silica-oversaturated phases consisting of 2 types: quartz-amphibole syenite and granite. The amphibole-pyroxene syenites at the Port Coldwell alkalic complex locally approach granite in composition and form large bodies. The quartz is interstitial to the feldspars and mafic minerals and is paragenetically late.

Granite occurs as irregular leucocratic dike-like bodies and narrow aplite veins. The Port Coldwell alkalic complex contains the most extensive development of these late granitic phases within an Ontario alkalic complex (Sage 1986, unpublished data). At the Port Coldwell alkalic complex, late pegmatites contain metre-scale cavities lined with euhedral quartz, feldspar and amphibole.

The Port Coldwell alkalic complex is cut by numerous north-trending lamprophyre dikes, which based on their trend, may be related to the TSTZ, but are not clearly related to the complex. Aubut (1977) classified a number of the dikes petrographically into the following groups: 1) fine-grained carbonate and zeolite-bearing ocellar lamprophyres; 2) aluminum-chromium augite-phyric lamprophyres; 3) fine-grained sericitized plagioclase lamprophyres with ocelli of carbonate and zeolite; and 4) quartz-phyric lamprophyres with minor clinopyroxene phenocrysts and small carbonate ocelli. The most common dikes are the quartz-phyric lamprophyres. The sericitized plagioclase lamprophyres cut the quartz-phyric lamprophyres with aluminum-chromium augite phenocrysts (Aubut 1977).

LACKNER LAKE

The Lackner Lake Alkalic Complex (Figure 18.10) occurs in the southern part of the KSZ. It is a ring complex, emplaced into felsic gneisses of the KSZ, and is composed of partial rings of ijolite, malignite, and nepheline syenite. Sheet-like bodies of carbonatite are conformable with the arcuate nature of the ijolite and occur within the ijolite. Dike-like bodies of titaniferous magnetite-apatite rock cut both the nepheline syenite and the mafic silicate phases. Two Rb-Sr whole rock ages from the complex have been determined. The ages are 1138±29 and 1092±21 Ma (Bell and Blenkinsop 1980; *see* Table 18.2).

The Lackner Lake Alkalic Complex contains elongated xenoliths of schist and gneiss, and trachytoidal planar structures oriented to conform with the outline of the intrusion (Sage 1988c). The arcuate sheet of malignite and ijolite within the complex is emplaced into a ring fracture in the syenite. The malignite-ijolite locally remobilized the nepheline syenite and has developed a zone of nepheline syenite hornfels (Sage 1988c).

CLAY–HOWELLS

The Clay–Howells Alkalic Complex is situated in the KSZ and is a good example of a large stock-like alkalic intrusion (Figure 18.11); however, it has not been well studied. The high-grade gneisses of the KSZ display little disruption, and xenolithic blocks of gneiss retain, within the intrusion, attitudes similar to that found within the enclosing gneisses from which they were derived. Emplacement was therefore most likely passive. Considering the relatively large surface area of the stock-like body, its most likely form appears to be a mushroom-shaped cross section representing the expanded top of a dike. This stock may be a relatively flat sheet fed by 1 or more dikes emplaced within faults associated with the KSZ. A septum of gneissic rocks striking south along the Mattagami and Kapuskasing rivers may be the result of ring fracturing and caldera subsidence of the roof rocks above the intrusion.

The complex is composed dominantly of a relatively homogeneous pyroxene syenite, which, on the basis of interpretation of the closure of isomagnetic contours, may be of a composite nature. A dike-like body of magnetite-bearing carbonatite intrudes the syenite in the southeast corner of the syenite complex. The carbonatite is enveloped in an alkalic mafic syenite that is possibly a product of fenitization related to the carbonatite. Rubidium-strontium whole rock ages on the syenitic rocks are 1076±17 Ma and 1072±16 Ma (Bell et al. 1979; Bell and Blenkinsop 1980; *see* Table 18.2).

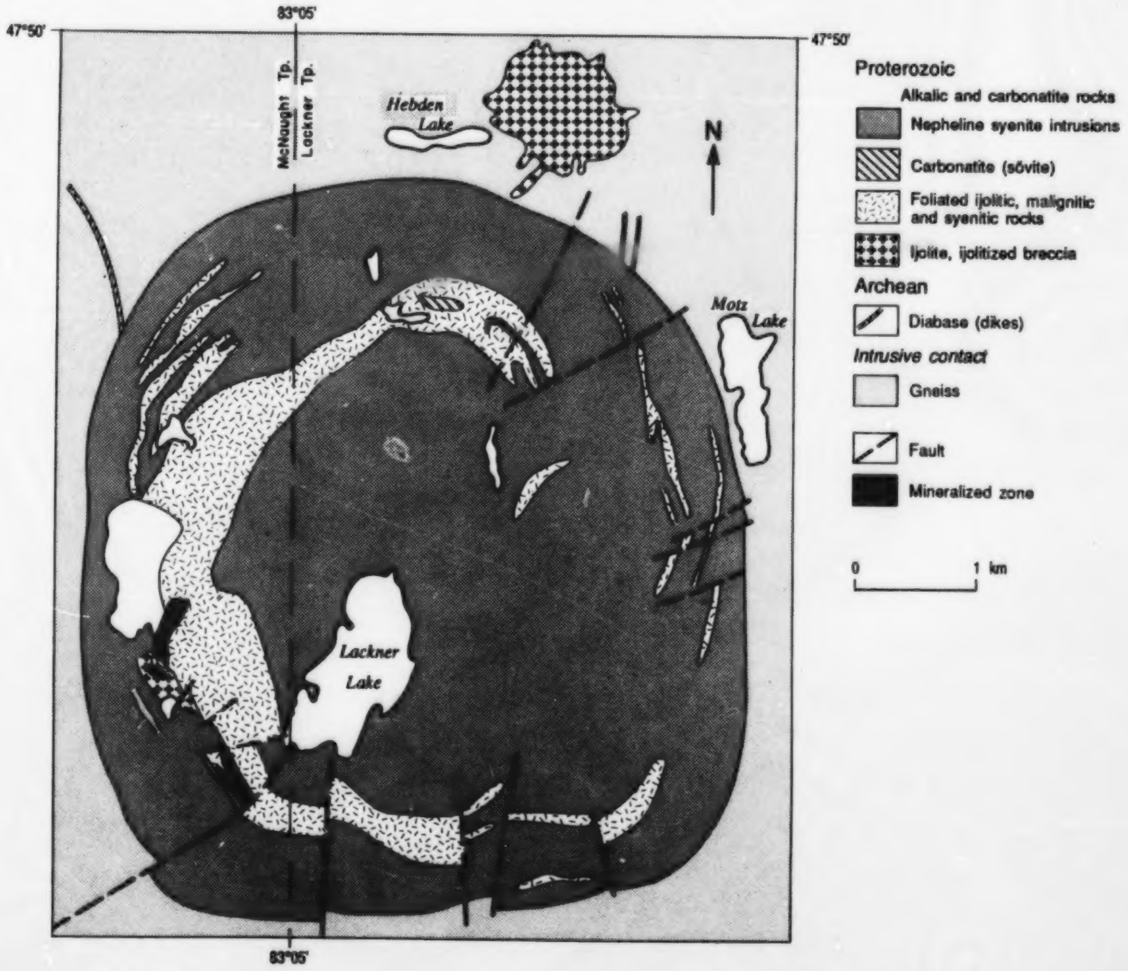
Carbonatite Complexes

Within Ontario, carbonatite complexes consist of 4 major rock types: sövite, silicocarbonatite, ijolite and pyroxenite. The Ontario carbonatite complexes display carbonatite-rich and/or carbonate-silicate rock types. The Firesand River Carbonatite Complex is largely carbonate, the Prairie Lake Carbonatite Complex displays a subequal development of carbonate and silicate units and the Seabrook Lake Carbonatite Complex is dominated by silicate rocks (Sage 1987a, 1988a, 1988b). The carbonatite complexes are enveloped by altered wall rocks referred to as fenites which exhibit desilicification and addition of iron, sodium, potassium and volatiles. The light REEs, LIL and other incompatible elements are noticeably concentrated in the altered rocks. Deuteric and metasomatic alteration within the intrusions has produced biotite-rich rocks, referred to as biotitite or glimmerite.

Carbonatites are enveloped in a fenitic aureole which in close proximity to the carbonatite is characterized by the pervasive development of green sodium-iron amphiboles and pyroxenes and a bulk desilication reaction. With increasing distance from the carbonatite, metasomatic alteration becomes controlled by fractures along which biotite and rusty weathering carbonate occur. Fenitic rocks enclosing Ontario carbonatites are commonly poorly exposed and have not been subjected to detailed study. Siemiatkowska and Martin (1975; *see* Figure 18.1), studied fenites near Sudbury which lack associated magmatic units, and classified these fenites as ultrasodic. These rocks are the product of desilication accompanied by the development of riebeckite (*see* Table 18.1), aegirine, albite and minor microcline feldspar (Siemiatkowska and Martin 1975).

Baldock (1973) and Vartianen and Woolley (1976) correlated widespread potassium fenitization with high-level emplacement and more restricted sodium fenitization with deep structural levels (Figure 18.12 and 18.13). Carbonatite magmas are vertically zoned with respect to mineralogy (LeBas 1981). The magmas have carbonate-rich phases at high structural levels, and silicate-rich phases at deeper levels (*see* Figure 18.12). With the exception of the diatreme structures in northeastern Lake Superior (Sage 1982), sodium fenitization is characteristic of Ontario carbonatites. The fenitization is dominantly sodium-iron, proximal to carbonatites and represented by green sodium-iron amphiboles and pyroxenes. The fenite aureole generally consists of an outer metasomatic halo represented by biotite and an inner zone composed of sodium-iron amphibole or clinopyroxene. The Prairie Lake and Seabrook Lake carbonatite complexes have relatively limited zones of sodium-iron fenitization and large volumes of mafic silicate units (Sage 1987a, 1988a; *see* Figure 18.12); therefore, these 2 carbonatites may represent relatively deep structural levels. The Firesand River Carbonatite Complex is dominated by carbonate-rich rocks, has a halo of fenitization up to 1.6 km wide, displays miarolitic cavities and is thus, most likely exposed at a high structural level (Sage 1988b).

The dominant carbonate mineral within Ontario carbonatites is calcite, with subsidiary dolomite. Sövite, a rock containing 50% or more calcite with subordinate dolomite, is a major lithologic unit in all Ontario carbonatite complexes. Sövite contains a wide variety of accessory minerals



83°05'
47°50'
47°50'
McNaught Tp.
Lackner Tp.
Hebden Lake
N
Motz Lake
Lackner Lake
83°05'
Proterozoic
Alkalic and carbonatite rocks
Nepheline syenite intrusions
Carbonatite (sövite)
Foliated ijolitic, malignitic and syenitic rocks
Ijolite, ijolitized breccia
Archean
Diabase (dikes)
Intrusive contact
Gneiss
Fault
Mineralized zone
0
1 km

Figure 18.10. Schematic sketch map of the geology of the Lackner Lake Alkalic Complex (Sage 1988c).

including olivine, clinohumite, sodic amphibole, phlogopite, biotite, apatite, magnetite, perovskite, pyrochlore and sodic clinopyroxene (*see* Table 18.1). High-iron carbonate carbonatite may intrude sövite and ferruginous rauhaugite is present in the Firesand River Carbonatite Complex (*see* Table 18.1). The general order of emplacement of carbonatite phases is sövite, rauhaugite and iron carbonate.

Silicocarbonatite is abundant in nearly all carbonatites. Its mineralogy is similar to sövite, but carbonate content varies from 10 to 50% . The rock type may originate from: 1) a cumulate phase in sövites; 2) a product of reaction of a carbonatite magma with an early phase or xenoliths; or 3) an altered and dismembered mafic dike.

Ijolitic and related melteigites and urtites are the most common silicate units associated with carbonatite complexes. Melteigite is usually formed by cumulate processes and urtite is usually a coarse-grained nepheline-rich pegmatitic phase. Clinopyroxenes in ijolites at Seabrook Lake are salite (variety of diopside) with aegirine-augite rims (Mitchell 1972) and at Prairie Lake, the clinopyroxenes are salite (Melnik 1984). The geology of a few selected Ontario carbonatites is given in the following section.

CARGILL TOWNSHIP

The Cargill Township Carbonatite Complex (Sage 1986) occurs in the central part of the KSZ. The complex consists of an outer rim of pyroxenite enclosing a carbonatite core (Figure 18.14). The core is composed of silicocarbonatite and sövite. Dolomitic and sideritic phases are likely present (Sandvik and Erdosh 1977) but have not been clearly delineated. The intrusion has been dated by U-Pb techniques as 1906±30 Ma (Kwon 1986; *see* Table 18.2). Karst weathering of the carbonatite in Cretaceous time has produced an apatite-rich residuum estimated at 62.5 million tons grading 19.6% P_2O_5 (Sandvik and Erdosh 1977). This residuum also contains rare earth element concentrations. The carbonatite is transected by a northeast-trending fault creating a dumbbell-shaped aeromagnetic pattern caused by offset of the generally circular isomagnetic contours (*see* Figure 18.14). The complex underwent postemplacement deformation and recrystallization which is evident by the development of cataclastic rock textures.

Based on a detailed study of the mineralogy of the Cargill Township Carbonatite Complex, Twyman (1983) proposed that the parental magma was a carbonated mafic alkaline magma that underwent immiscibility. This gave rise to an olivine sövite magma which subsequently differentiated into a natrocarbonatite magma (*see* Table 18.1) via crystal fractionation. The Cargill Township Carbonatite Complex contains olivine (Fo_{82}), in masses up to 5 cm in diameter, which is commonly replaced by high-titanium clinohumite. This replacement likely represents late reaction of carbonatite magma with early formed olivine (Sage 1988h).

The mica in sövite from Ontario carbonatites often displays dark reddish-brown rims and pale yellow-brown

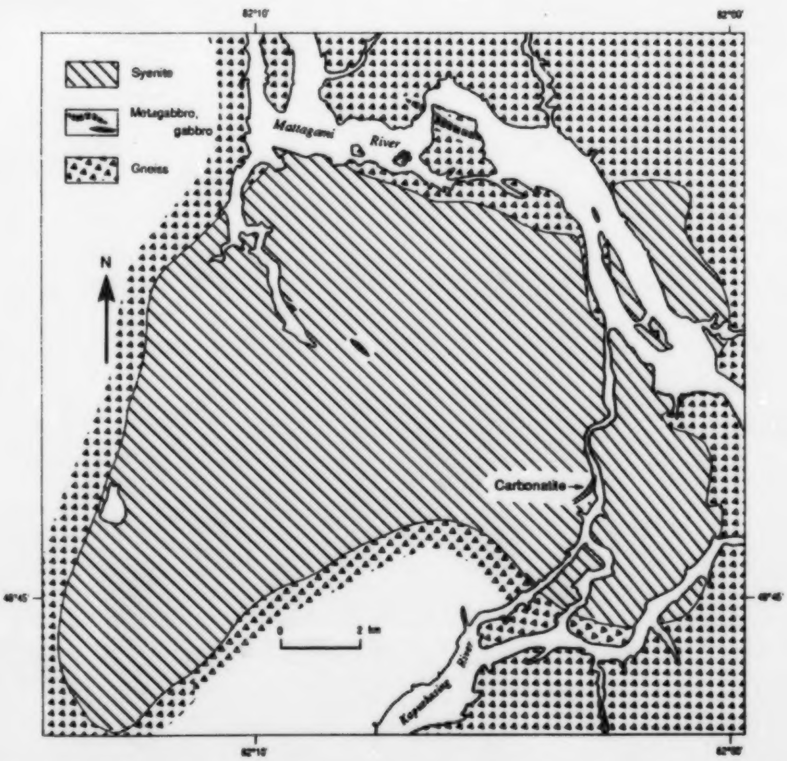


Figure 18.11. Schematic sketch of the geology of the Clay–Howells Alkalic Complex (Sage 1988g).

cores. The best characterized material is from the Cargill Township Carbonatite Complex where Gasparrini et al. (1971) found iron enrichment towards grain margins. This suggests that the early phlogopitic cores reacted with late-stage iron-rich magmatic fluids to form biotitic rims (Sage 1988h).

Pyroxenite is not a common phase in Ontario carbonatites. It is restricted to three 1.9 billion-year-old carbonatites at the north end of the KSZ; Cargill Township, Argor and Goldray complexes (Sage 1988h, 1988i, 1988k). The clinopyroxene has been well characterized at the Cargill Township Carbonatite Complex and ranges from diopside to augite in composition. The clinopyroxene is associated with subordinate amphibole of pargasite or hastingsite composition (*see* Table 18.1; Allen 1972; Twyman 1983; Sage 1988h). Carbonatite intrudes the pyroxenite and has reacted with the pyroxenite to form phlogopite (Gittins et al. 1975).

FIRESAND RIVER

The Firesand River Carbonatite Complex intrudes Archean rocks of the Wawa Subprovince of the Superior Province and is emplaced along faults associated with the KSZ. Two K-Ar ages for this intrusion are 1008 and 1087 Ma (Gittins et al. 1967; *see* Table 18.2). The complex consists of sövite, silicocarbonatite, ijolite and ferruginous rauhaugite (Figure 18.15). The carbonatite is zoned with sövite and silicocarbonatite at the periphery of the intrusion and rauhaugite in the core. The ijolite does not form a distinct unit, but occurs as blocks metasomatized by fenitizing fluids in the sövite rim of the complex. The sövite and rauhaugite appear gradational into each other. The effects of fenitization associated with the complex have been observed by the author along joint surfaces as far as 1.6 km from the complex. The fenitization effects, close to the carbonatite, are indicated by the presence of dark blue-green amphiboles replacing quartz in the wall rocks and along fractures. Biotite and ferruginous carbonate are commonly present along fractures in the wall rock more distant from the complex. Lamprophyre dikes with associated iron carbonate and sodium amphibole alteration, characteristic of fenitization, are common in the area near the Firesand River Carbonatite Complex.

The distinctive ferruginous dolomite core locally displays small miarolitic cavities up to 2 cm in diameter, lined with carbonate, quartz and barite, and fractures filled with quartz and barite (Sage 1988b). Skeletal olivine intergrown with carbonate found in the Firesand River Carbonatite Complex suggests that quenching of the carbonatite magma

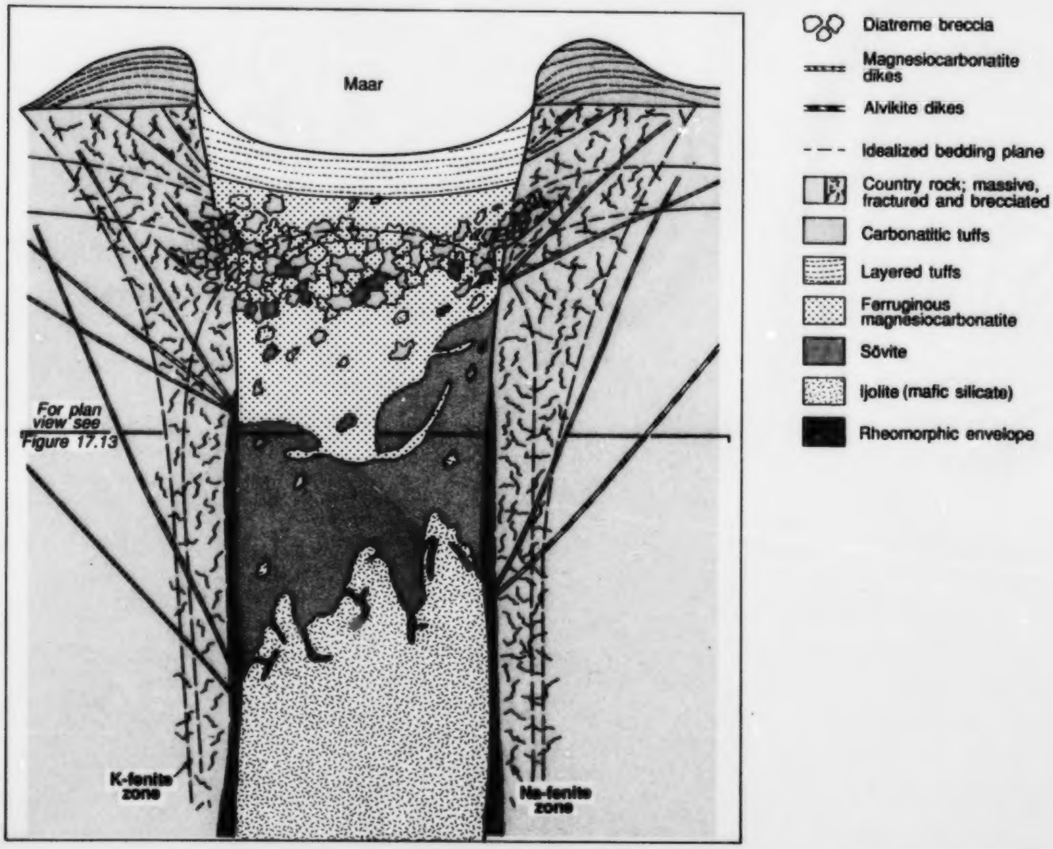


Figure 18.12. Idealized vertical cross section of a carbonatite intrusion. The model is based on mapping of Ontario carbonatite and diatreme structures (Sage 1986). The sodium-rich fenite zone undergoes desilication whereas the potassium-rich fenite zone undergoes only moderate desilication.

occurred (Sage 1988b). This unusual olivine-carbonate texture has also been found in the Schryburt Lake and Valentine Township Carbonatite complexes (Sage 1988i, 1988j).

Amphibole is ubiquitous and generally acicular consisting of richterite, magnesio-arfvedsonite (magnesium-sodium-rich amphibole), arfvedsonite (sodium-rich amphibole) and ferrorichterite (sodium-iron-rich amphibole), at the Firesand River, Argor and Cargill Township Carbonatite complexes (Teal 1979; Sage 1988b, 1988h, 1988k). Teal (1979) found clinopyroxene in sövite at the Firesand River Carbonatite Complex to vary from aegirine to diopside in composition.

PRAIRIE LAKE

The Prairie Lake Carbonatite Complex intrudes Archean rocks of the Wawa Subprovince. The complex lies along fractures associated with the TSTZ and has a Rb-Sr whole rock age of 1165±30 Ma (Pollock 1987; *see* Table 18.2). The complex forms a high hill on the west side of Prairie Lake, but deep weathering has inhibited preservation of fresh rock at surface. The complex is composed of ijolite, wollastonite ijolite, sövite and silicocarbonatite (Figure 18.16; *see* Table 18.1). The complex is enveloped by a very poorly exposed fenitized granitic terrane. A linear aeromagnetic pattern suggests that a lineament, possibly parallel to the lithologic trends, connects the Prairie Lake Carbonatite Complex with the Killala Lake Alkalic Complex. The carbonatite occurs at the intersection of the aeromagnetically inferred lineament and a topographic lineament that extends north from the area of the Slate Islands.

Limited microprobe study of the Prairie Lake minerals has been completed by D. Watkinson, Carleton University and compiled by Sage (1987a). Pyroxene and nepheline within the ijolitic rocks are highly variable in composition. The pyroxenes vary from aegirine-augite to salite in composition and the nepheline varies between high-nepheline, $(Na,K)\ AlSiO_4$, and high-kalsilite, $KAlSiO_4$, compositions (Sage 1987a). The accessory garnet found in the ijolite rock is unusual in that it contains up to 5% Zr (Sage 1987a).

The sövite or carbonatite-rich phase is dominantly calcite with intergrown dolomite containing accessory apatite, amphibole, biotite-phlogopite, magnetite, perovskite and olivine (Sage 1987a). A minor ferruginous ankerite unit occurs in the centre of the complex which contains uranium-niobium mineralization in the form of pyrochlore and betafite (Sage 1987a).

SEABROOK LAKE

The Seabrook Lake Carbonatite Complex intrudes Archean rocks of the Abitibi Subprovince of the Superior Province. The complex occurs at the intersection of several regional lineaments and is the southernmost alkalic-carbonatite complex associated with the KSZ. The complex is composed of ijolite, sövite, and silicocarbonatite enveloped by fenitized breccia (Figure 18.17). The complex has been

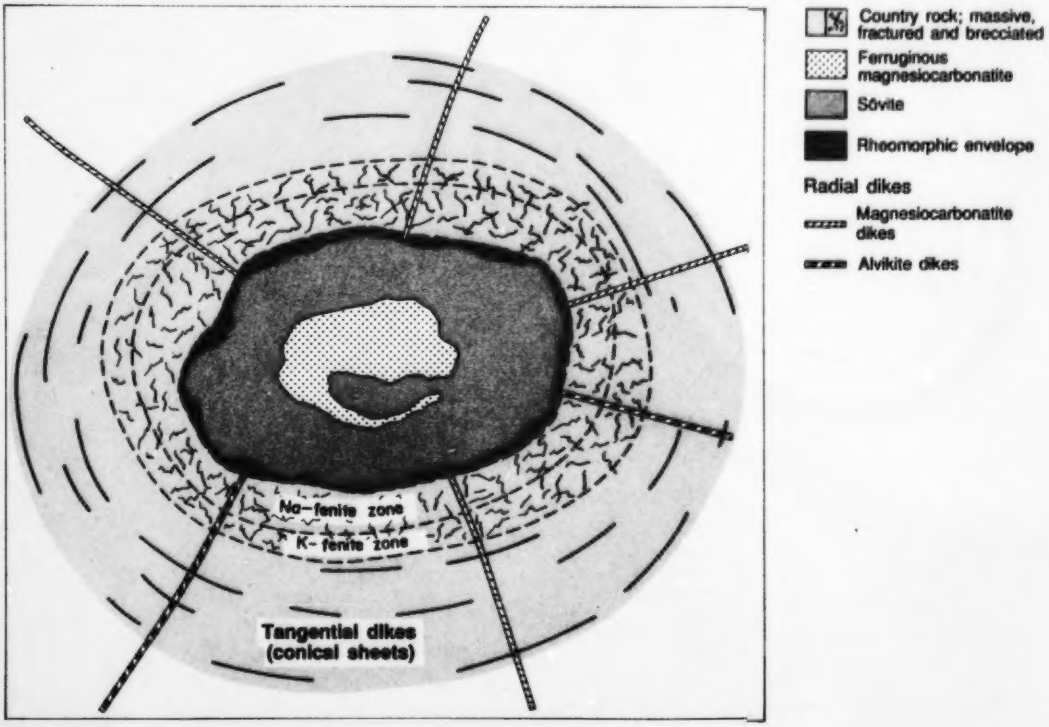

Country rock; massive, fractured and brecciated
Ferruginous magnesiocarbonatite
Sövite
Rhoomorphic envelope
Radial dikes
Magnesiocarbonatite dikes
Alvikite dikes
Na-fenite zone
K-fenite zone
Tangential dikes (conical sheets)

Figure 18.13. Idealized plan view of a simple ring complex based on mapping of Ontario carbonatite and diatreme structures (Sage 1986). The rheomorphic envelope may not be present in all cases. Within the complex, sövite may contain altered mafic silicate blocks.

dated by K-Ar isotopic techniques on biotite as being approximately 1100 Ma (Gittins et al. 1967, *see* Table 18.2).

The Seabrook Lake Carbonatite Complex is dominated by rocks of the ijolite suite; clinopyroxene ranges in composition from aegirine-augite at the rim to salite at the core (Mitchell 1972). The nepheline contains up to 25% kalsilite molecule and is iron-rich (Mitchell 1972; Parbery 1984).

Kimberlite

Kimberlite is very rare in Ontario. The few K-Ar isotopic ages available for this group of rocks are from northeastern Ontario and indicate that they have a Jurassic age. Kimberlite occurs as dikes (hypabyssal facies) and breccias (diatreme facies). Narrow dikes and plugs of kimberlite have been found by drilling targets identified through detailed ground magnetic surveys and through indicator mineral trails of pyrope garnet and chrome diopside.

Over the years, the indicator minerals pyrope, magnesian ilmenite and chrome-diopside have been found in the Hudson Bay Lowlands and the Kirkland Lake area (Brown et al. 1967; Lee 1968; Satterly 1971; Brummer 1978). These finds have stimulated numerous exploration efforts for diamond which continue to the present.

The Lake Timiskaming Fault System (*see* Figures 18.2 and 18.18) hosts the Kirkland Lake kimberlites and the greatest number of known kimberlites in Ontario. These kimberlites occur along the flanks of the structural zone and its extension to Kirkland Lake (Figure 18.18). Satterly (1949) first reported the occurrence of a 15 cm wide

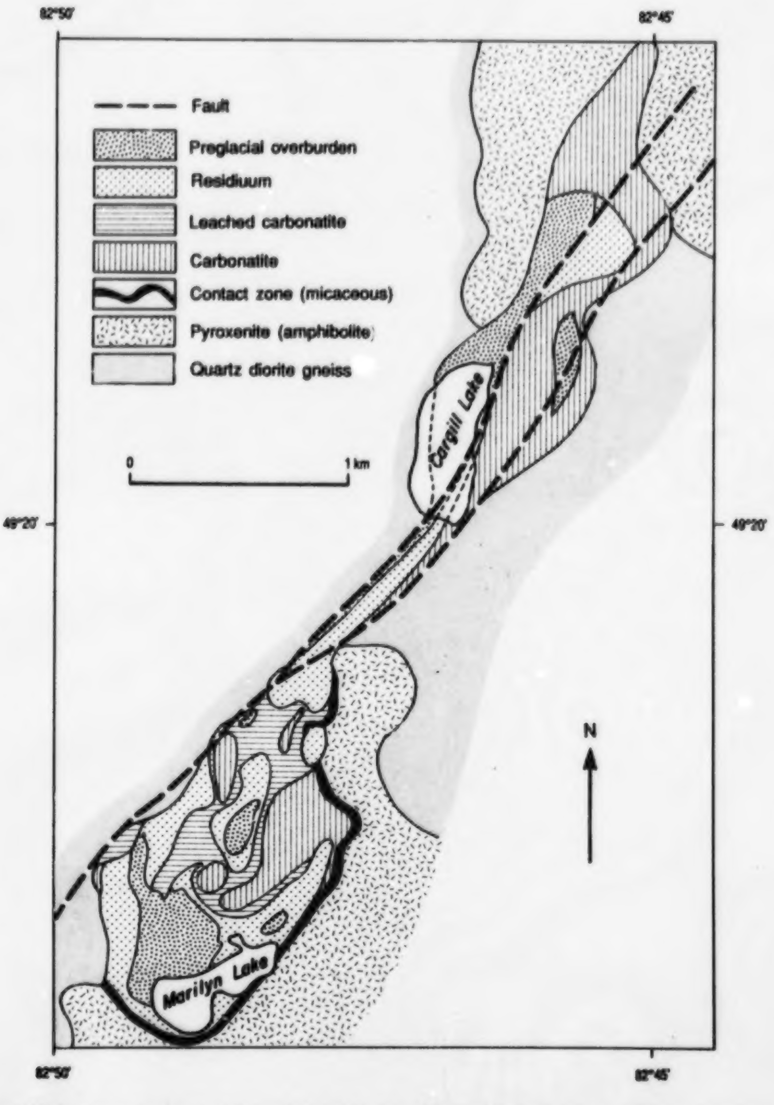

82°50'
82°45'
Fault
Preglacial overburden
Residuum
Leached carbonatite
Carbonatite
Contact zone (micaceous)
Pyroxenite (amphibolite)
Quartz diorite gneiss
0
1 km
49°20'
49°20'
Cargill Lake
N
Marilyn Lake
82°50'
82°45'

Figure 18.14. Schematic sketch of the geology of the Cargill Township Carbonatite Complex (Sage 1988b).

kimberlite dike in this area, but more recently, several references to kimberlite along the Lake Timiskaming Fault System between Kirkland Lake and Lake Nipissing have been made (Lovell et al. 1987; Owsiacki 1987; Grabowski et al. 1988).

Of the Kirkland Lake area kimberlites, the best documented are the hypabyssal dikes found in the Upper Canada Mine (Lee and Lawrence 1968; Shirley 1972). This kimberlite contains serpentinized olivine, phlogopite, chromite, titaniferous magnetite, perovskite, pyrope garnet, apatite, serpentine and carbonate. The carbonate is calcite and shortite (Shirley 1972; Watkinson and Chao 1973). This Upper Canada Mine kimberlite has been dated by K-Ar isotopic techniques at 151±8 Ma (Lee and Lawrence 1968), and at 158±2 Ma by U-Pb techniques (Heaman 1989). Near Kirkland Lake, float kimberlite boulders with limestone xenoliths contain Middle Ordovician to Middle Devonian conodonts (Grabowski et al. 1988). Microdiamonds have been found in the Alfie Creek No. 1 kimberlite pipe northeast of Kirkland Lake (Lovell et al. 1987).

In the Hudson Bay Lowlands, numerous alkalic ultrabasic diatremes intrude the Paleozoic rocks (Janse et al. 1986; Figure 18.19). These diatremes have been classified by exploration companies as ultramafic tuff breccia, alnoite (*see* Table 18.1) and carbonatite (Janse et al. 1986). True kimberlite has not been found in the Hudson Bay Lowland, even though K-Ar isotopic age dates on the alnoites of 152±8 Ma and 180±9 Ma (Janse et al. 1986) indicate that at least some of these bodies are contemporary with kimberlites in the Upper Canada Mine.

Outside the area of the Hudson Bay Lowland and the Lake Timiskaming Fault System, reference to the presence of kimberlite or kimberlitic rocks is rare.

Within the central part of the KSZ, Watson et al. (1978) described a 0.4 m wide kimberlite dike. Olivine melilite at Coral Rapids (Butler et al. 1988), also within the KSZ, has been described by Brown et al. (1967) and Bennett et al. (1967) as "kimberlitic".

Barnett et al. (1984) described Jurassic kimberlitic dikes near Belleville, at the western end of the St. Lawrence Graben System. The dikes referred to as the Varty and Picton dikes, are small and may be related to the extension of the northeast-trending Clarendon–Linden Fault outlined in adjoining New York State. The Varty dike has been dated at 176±3 Ma using U-Pb techniques (L.M. Heaman, Royal Ontario Museum, personal communication, 1990).

ECONOMIC GEOLOGY

Ring complexes, such as the ones at Port Coldwell and Killala Lake, with a gabbro component, have significant economic potential due to the presence of disseminated

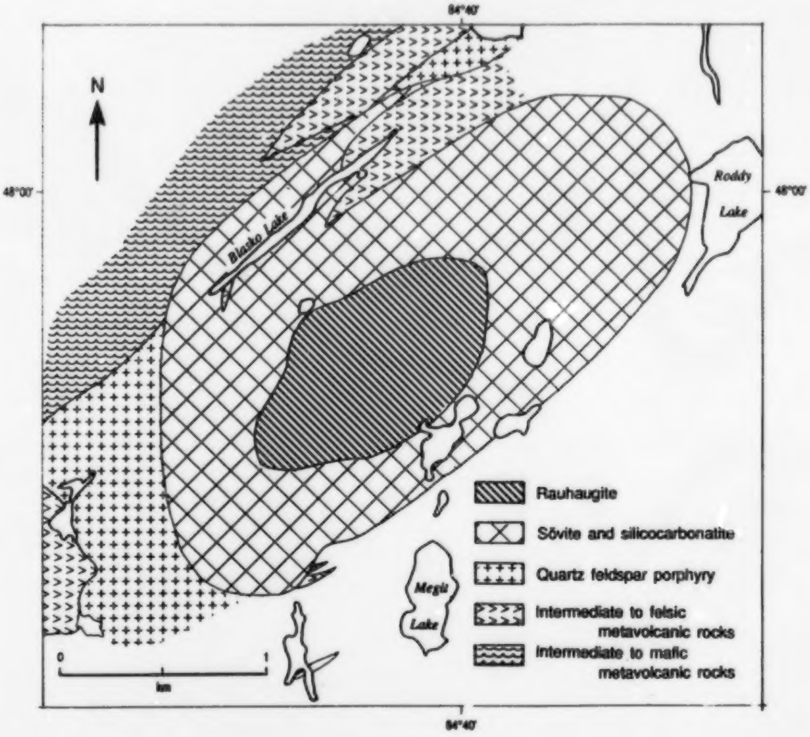

84°40'
N
48°00'
48°00'
Blasko Lake
Roddy Lake
Megit Lake
Rauhaugite
Sövite and silicocarbonatite
Quartz feldspar porphyry
Intermediate to felsic metavolcanic rocks
Intermediate to mafic metavolcanic rocks
0
1
km
84°40'

Figure 18.15. Schematic sketch of the geology of the Firesand River Carbonatite Complex (Sage 1988b).

copper and nickel sulphides with platinum group element (PGE) concentrations. In the Port Coldwell alkalic complex, the outer gabbro ring of the oldest of 3 ring complexes hosts in excess of 60 million tons grading less than 0.5% combined Cu and Ni with PGE values (Sage 1986). The syenitic rocks in ring complexes also have the potential to supply building stone.

Nearly all Ontario carbonatites contain phosphate and niobium. The Argor Carbonatite Complex contains the richest known niobium-bearing primary deposit in Ontario, 62 million tons grading 0.52% Nb_2O_5 outlined by diamond drilling. Rare earth elements and uranium are enriched within Ontario carbonatites, but large reserves have not been outlined. Complexes of the same age and similar geology vary widely in their economic potential, for example, the niobium-rich Argor Carbonatite Complex contrasts with the nearby niobium-poor Goldray Carbonatite Complex (Sage 1988k, 1988l) of similiar age and geology. Crystal fractionation concentrates niobium in late phases (Twyman 1983) and the least-differentiated olivine-rich carbonatite, such as the Cargill Township Carbonatite Complex, contains less mineralization than the relatively olivine-poor differentiated complexes, such as the one at Argor. Olivine is present only as a very minor phase in the Firesand River, Schryburt Lake and possibly Prairie Lake carbonatite complexes (Sage 1987a, 1988b, 1988i) all of which contain widespread low-grade niobium mineralization. The differentiation model of Twyman (1983) is a promising exploration guide, but it needs further evaluation.

Deep weathering of carbonatites in Ontario has produced concentrations of accessory or secondary minerals rich in niobium, REEs, uranium and phosphorous. This

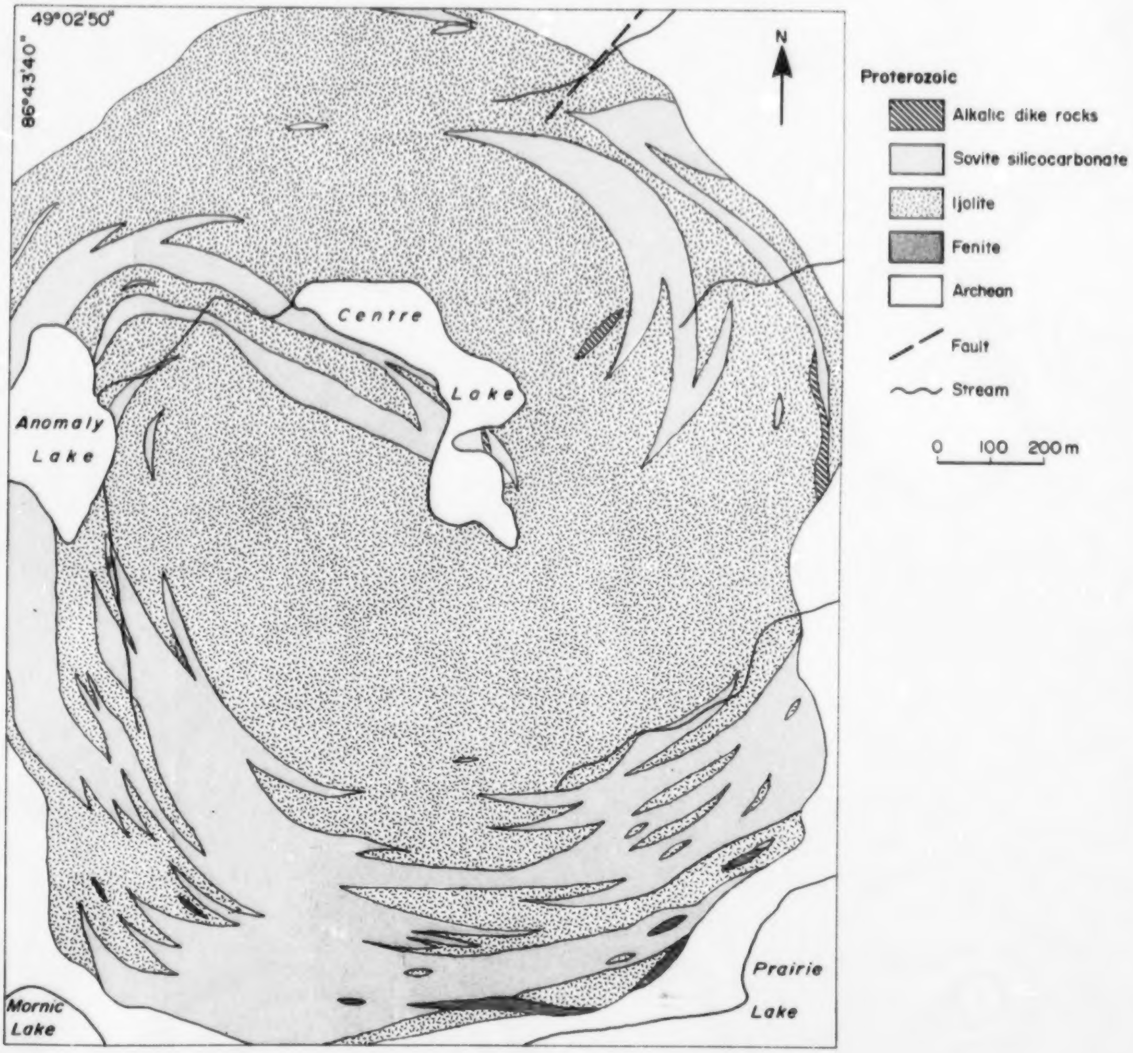


Figure 18.16. Schematic geological map of the Prairie Lake Carbonatite Complex (Sage 1987a).

weathering process took place in the Cretaceous and carbonatites developed karst-like structures (Sage 1986, 1988h). Mapping, drilling and geophysics indicate that those complexes developing karst features are faulted. This faulting may be a necessary feature for the development of residual mineral deposits by providing channels for groundwater movement. Weathering develops concentrations of economically important elements that are up to 10 times the primary values found in unweathered carbonatite. The Cargill Township and Martison Carbonatite complexes are excellent examples of rich concentrations of phosphorous, niobium and REEs which are products of the deep weathering of a carbonatite complex (Sage 1986, 1988h).

SUMMARY

Proterozoic and younger alkalic-carbonatite complexes of Ontario are closely associated with regional crustal-scale structures. The peaks of alkalic activity are at 2.7 Ga, 1.9 Ga, 1.0 Ma, 570 Ma and 150 Ma. Crustal fractures controlling alkaline magmatism display no evidence of the migration of the intrusive centre with time along the structural trends. Isotope studies indicate that the alkaline magmas were generated from LIL-depleted upper mantle; however, LIL enrichment of this depleted mantle must occur in order to repeatedly form alkaline magmas over a long period of time. Alkalic intrusions of widely separated ages occur in

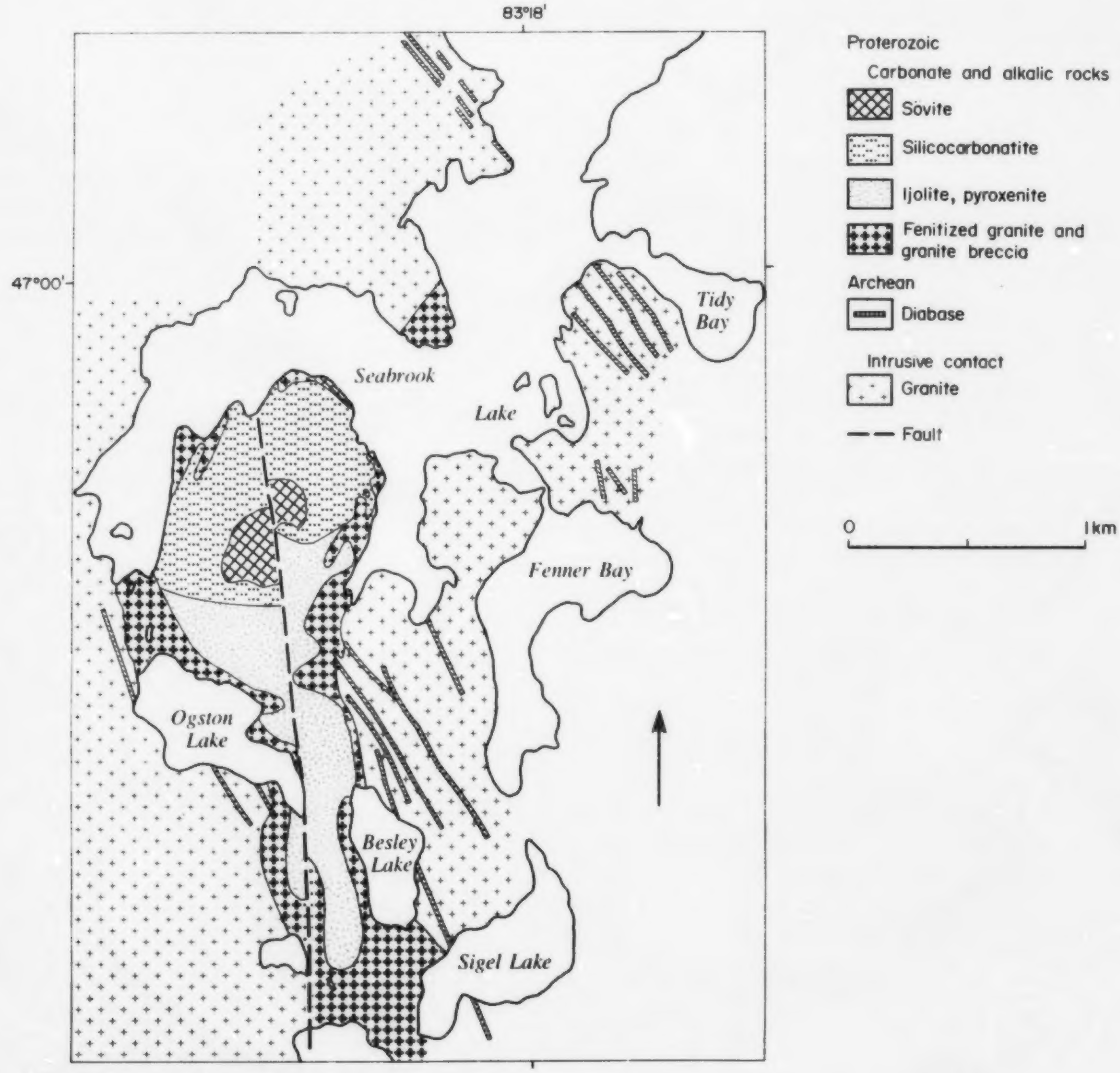

83°18'
47°00'
Seabrook
Lake
Tidy Bay
Fenner Bay
Ogston Lake
Besley Lake
Sigel Lake
Proterozoic
Carbonate and alkalic rocks
Sövite
Silicocarbonatite
Ijolite, pyroxenite
Fenitized granite and granite breccia
Archean
Diabase
Intrusive contact
Granite
Fault
0
1km

Figure 18.17. Schematic geological map of the Seabrook Lake Carbonatite Complex (Sage 1988a).

close proximity to each other, implying the process of mantle metasomatism has taken place to refertilize the mantle. The metasomatic process is perhaps controlled in part by the location of deep regional crustal fractures.

The alkalic complexes occur as relatively homogenous stocks representing the expanded tops of dikes or as ring complexes displaying ring fractures and cauldron collapse.

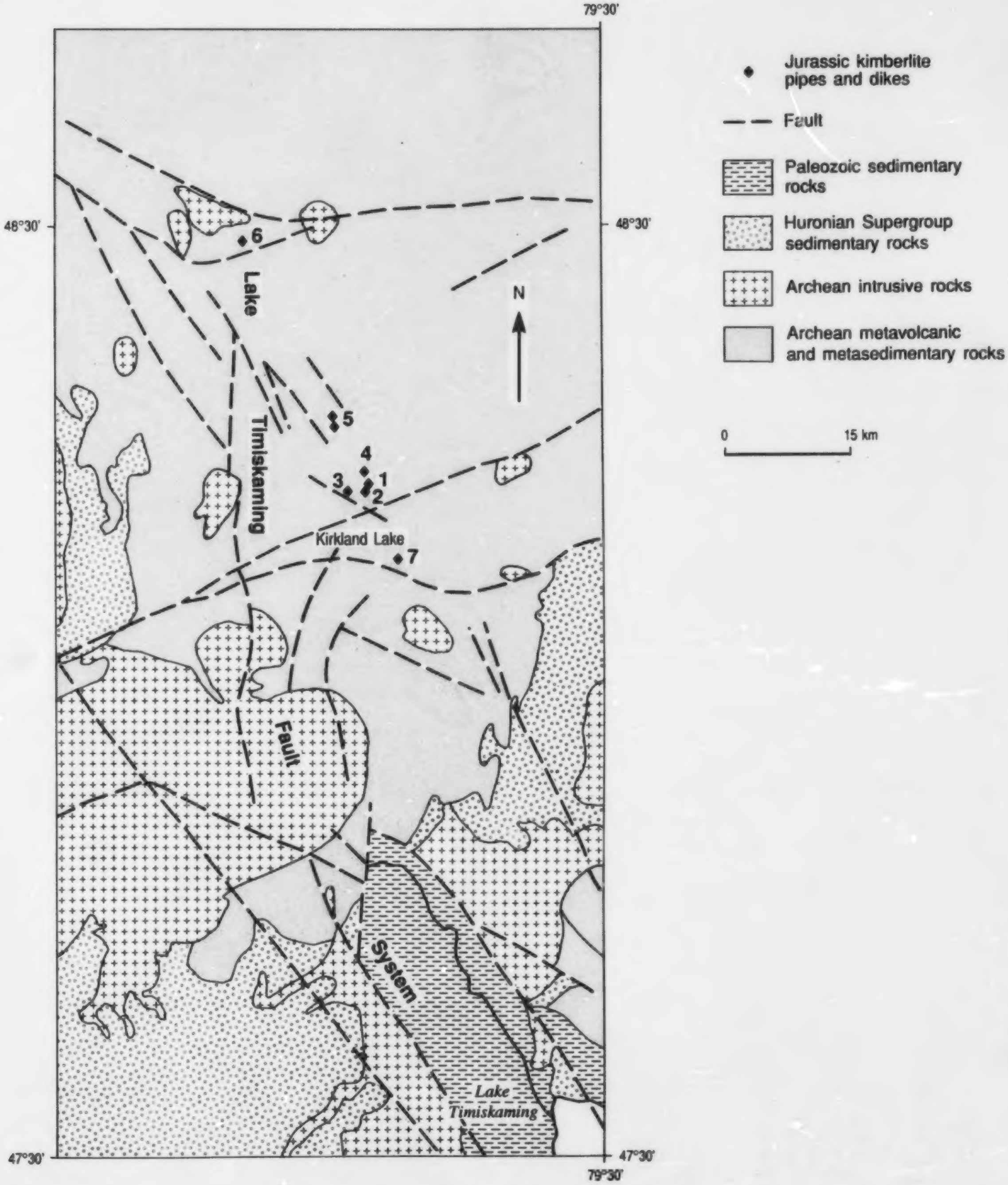

79°30'
48°30'
48°30'
47°30'
47°30'
79°30'
N
Lake Timiskaming Fault System
Kirkland Lake
Lake Timiskaming
1
2
3
4
5
6
7
Jurassic kimberlite pipes and dikes
Fault
Paleozoic sedimentary rocks
Huronian Supergroup sedimentary rocks
Archean intrusive rocks
Archean metavolcanic and metasedimentary rocks
0
15 km

Figure 18.18. Schematic sketch of the Kirkland Lake area indicating known kimberlite occurrences (*see* Figure 18.2 for regional structures). 1-Alfie Creek 1 pipe; 2-Alfie Creek 2 pipe; 3-Morrisette Creek pipe; 4-Nelson Lake pipe; 5-Nickila Lake pipe; 6-Satterly dike; 7-Upper Canada dike.

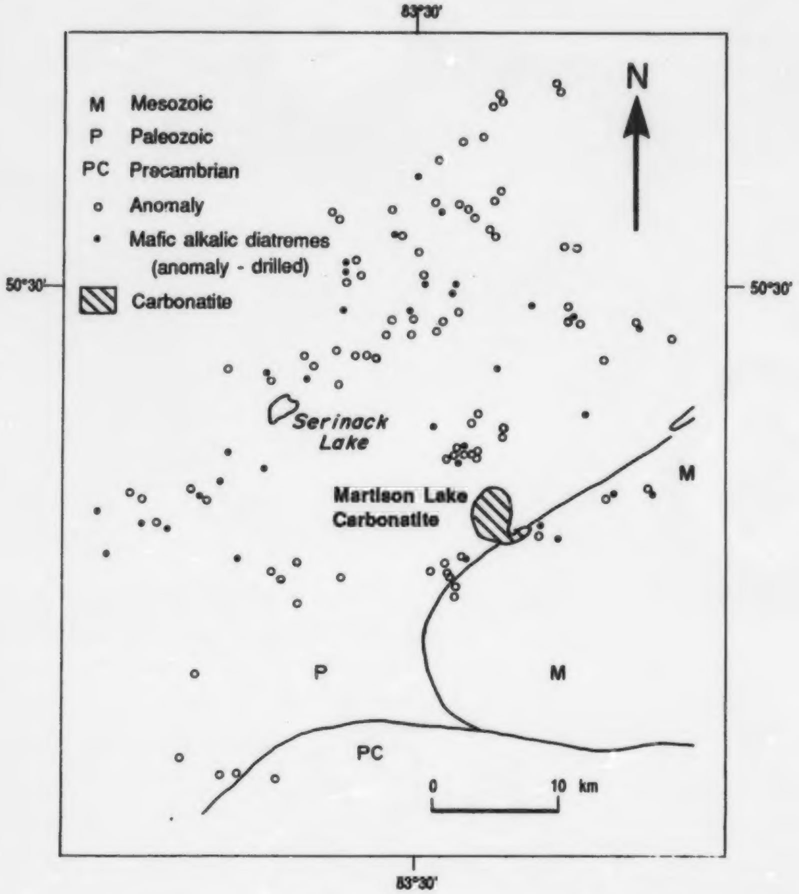


Figure 18.19. Schematic sketch of a part of the Hudson Bay Lowlands indicating numerous magnetic anomalies and those established as mafic alkalic diatremes (*see* Figure 18.1 for the location of the Martison Carbonatite Complex).

Carbonatites occur as small plug-like bodies with diameters of 3 to 4 km. Kimberlites are present as dikes and plugs.

Gabbroic phases of ring complexes are favourable hosts for copper, nickel and PGE mineralization. Carbonatites are enriched in phosphorus, niobium, REEs and uranium relative to other crustal rocks. These intrusive complexes offer a potential source for primary and secondary deposits of phosphorous, niobium and rare earths.

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CONVERSION FACTORS FOR MEASUREMENTS IN ONTARIO GEOLOGICAL SURVEY PUBLICATIONS

Conversion from SI to Imperial			**Conversion from Imperial to SI**		
SI Unit	*Multiplied by*	*Gives*	*Imperial Unit*	*Multiplied by*	*Gives*
			LENGTH		
1 mm	0.039 37	inches	1 inch	**25.4**	mm
1 cm	0.393 70	inches	1 inch	**2.54**	cm
1 m	3.280 84	feet	1 foot	**0.304 8**	m
1 m	0.049 709 7	chains	1 chain	20.116 8	m
1 km	0.621 371	miles (statute)	1 mile (statute)	**1.609 344**	km
			AREA		
1 cm^2	0.155 0	square inches	1 square inch	**6.451 6**	cm^2
1 m^2	10.763 9	square feet	1 square foot	**0.092 903 04**	m^2
1 km^2	0.386 10	square miles	1 square mile	2.589 988	km^2
1 ha	2.471 054	acres	1 acre	0.404 685 6	ha
			VOLUME		
1 cm^3	0.061 02	cubic inches	1 cubic inch	**16.387 064**	cm^3
1 m^3	35.314 7	cubic feet	1 cubic foot	0.028 316 85	m^3
1 m^3	1.308 0	cubic yards	1 cubic yard	0.764 555	m^3
			CAPACITY		
1 L	1.759 755	pints	1 pint	0.568 261	L
1 L	0.879 877	quarts	1 quart	1.136 522	L
1 L	0.219 969	gallons	1 gallon	**4.546 090**	L
			MASS		
1 g	0.035 273 96	ounces (avdp)	1 ounce (avdp)	28.349 523	g
1 g	0.032 150 75	ounces (troy)	1 ounce (troy)	**31.103 476 8**	g
1 kg	2.204 62	pounds (avdp)	1 pound (avdp)	**0.453 592 37**	kg
1 kg	0.001 102 3	tons (short)	1 ton (short)	**907.184 74**	kg
1 t	1.102 311	tons (short)	1 ton (short)	**0.907 184 74**	t
1 kg	0.000 984 21	tons (long)	1 ton (long)	**1016.046 908 8**	kg
1 t	0.984 206 5	tons (long)	1 ton (long)	**1.016 046 908 8**	t
			CONCENTRATION		
1 g/t	0.029 166 6	ounce (troy)/ ton (short)	1 ounce (troy)/ ton (short)	34.285 714 2	g/t
1 g/t	0.583 333 33	pennyweights/ ton (short)	1 pennyweight/ ton (short)	1.714 285 7	g/t

OTHER USEFUL CONVERSION FACTORS

	Multiplied by	
1 ounce (troy) per ton (short)	20.0	pennyweights per ton (short)
1 pennyweight per ton (short)	0.05	ounces (troy) per ton (short)

Note: Conversion factors which are in bold type are exact. The conversion factors have been taken from or have been derived from factors given in the Metric Practice Guide for the Canadian Mining and Metallurgical Industries, published by the Mining Association of Canada in co-operation with the Coal Association of Canada.